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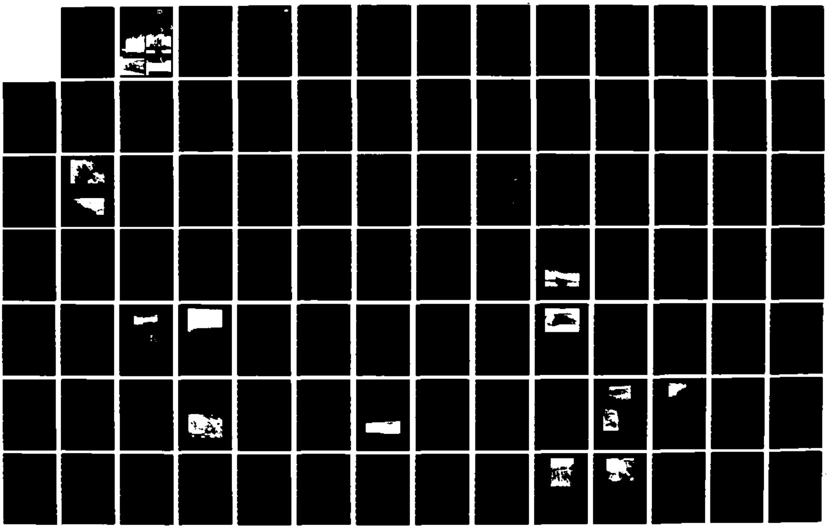

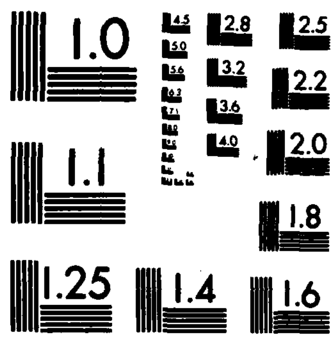

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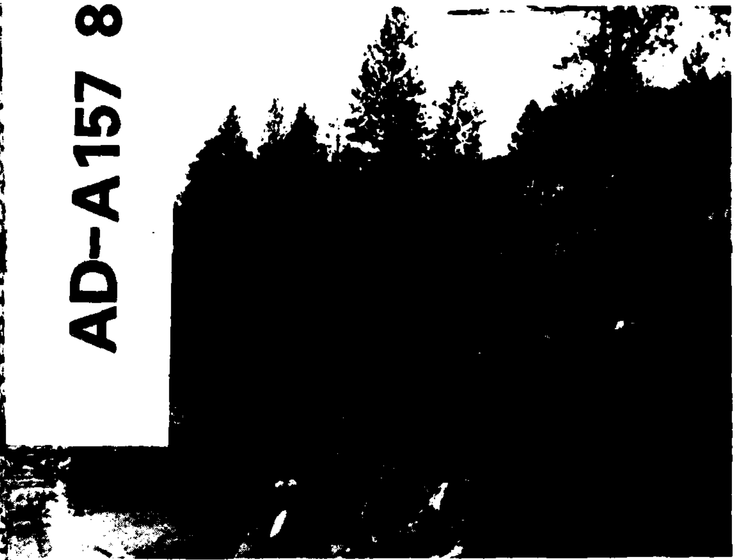
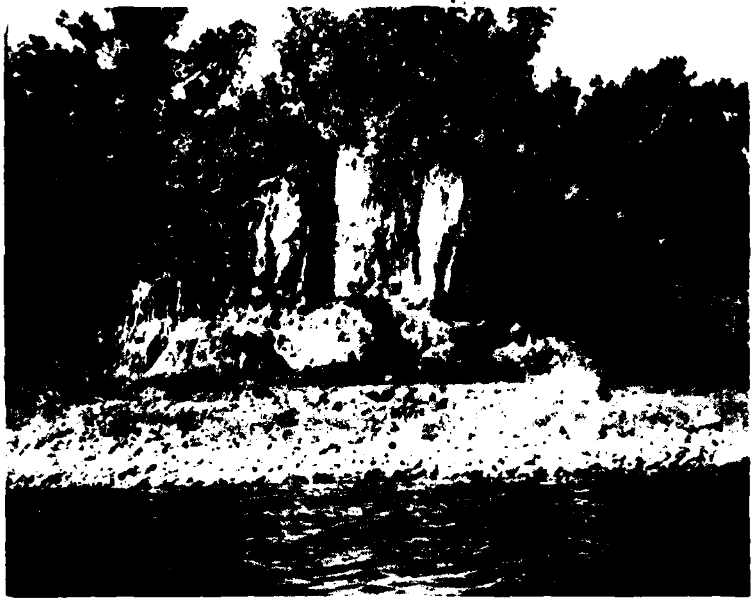
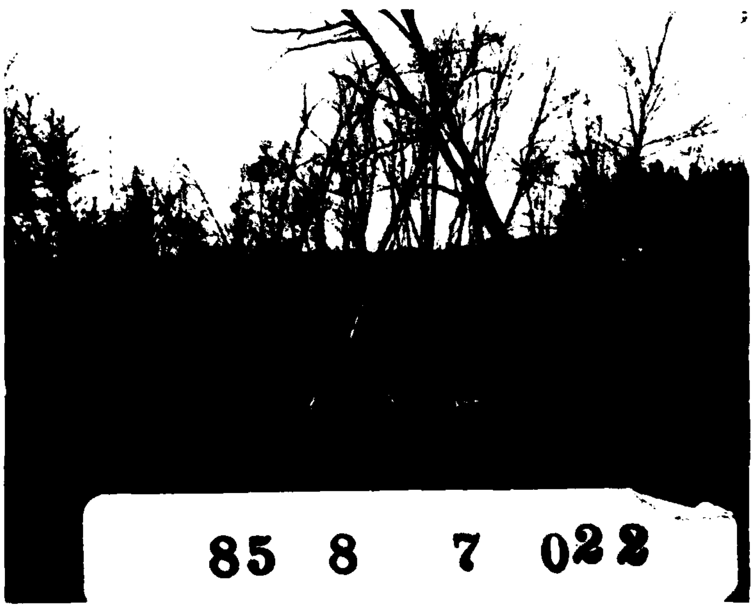
Erosion of northern reservoir shores
An analysis and application of pertinent literature







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Cover: Eroding and receding shore zones in northern reservoirs (clockwise from upper left): 1) Shore composed of interstratified glaciolacustrine and glaciofluvial sediments, Lake Koocanusa, Montana; 2) Beach exposed at a low water level, Orwell Lake, Minnesota, bluff height approximately 4 m; 3) Undermined and overturned blocks of sediment held together by vegetation in foreground, with slump blocks in background, Berlin Lake, Ohio; and 4) Large bluff composed of stratified glacial and glaciofluvial deposits on Lake Koocanusa behind Libby Dam, Montana.

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Erosion of northern reservoir shores

An analysis and application of pertinent literature

Daniel E. Lawson

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This monograph describes the current state of knowledge of northern reservoir shore erosion, primarily by examining the results of erosional studies on lakes, coasts and rivers. The major erosional processes of reservoir beaches and bluffs and their mechanics are discussed in detail. Thermal and physical parameters affecting the erodibility of shores, the environmental impacts of erosion, and the basic characteristics of the unique reservoir environment are reviewed. Current models of shore zone development are also presented. This literature analysis revealed that knowledge of erosion and recession in northern impound-

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ments is severely limited. Quantitative analyses of the processes of erosion and their relative importance, parameters determining the nature, rate and timing of erosion, and models to predict the erodibility of a shore for use in minimizing shoreline recession remain in need of basic field research.

PREFACE

This monograph was prepared by Dr. Daniel E. Lawson, Research Physical Scientist, of the Earth Sciences Branch, Research Division, U.S. Army Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire. This analysis was funded by the Office of the Chief of Engineers, Washington, D.C., as part of the Civil Works Project CWIS 31568, Erosion Potential of Inland Shorelines and Embankments in Regions Subject to Freezing and Thawing.

The intent of this monograph is to provide Corps personnel with a background in the environmental problem of erosion in northern reservoirs. In particular, it addresses the causes of erosion and shoreline recession, as well as the environmental impacts resulting from erosion. Various aspects of the unique reservoir environment are also discussed. It includes a review of the basic principles and mechanics of erosional processes and factors affecting them, and assesses the needs for further research. It is written such that individual sections are basically complete in themselves so that readers can select individual topics of interest without having to read the entire monograph. Some repetition will necessarily result from presenting the material in this way. Because of the complexity of this topic and the number of references related to it, some citations have undoubtedly been overlooked, but it is hoped that this review will provide sufficient background for an initial analysis of erosion problems as they arise, and act as guidance and stimulus for further investigations.

A contracted literature search prepared by D.B. Simons, M.A. Alawady and R.M. Li in 1978 and a contracted general literature review by C.A. Sternbach and G.M. Friedman in 1980 have contributed to this report. Observations at selected Corps reservoirs in northern regions and helpful discussions with Corps personnel during the period 1978 to 1984 were also incorporated into this monograph.

The author thanks CRREL librarians Nancy Liston, Elisabeth Smallidge, and Cynthia Whitney for their assistance in conducting literature searches and obtaining copies of papers and reports necessary for completing this work. He particularly thanks Nancy Richardson for typing the numerous drafts of this manuscript, Edmund Wright for his thoughtful editorial comments and review, and also Eleanor Huke for drafting figures used in this monograph. The author also thanks Lawrence Gatto, Michael Ferrick and James Wuebben for technically reviewing the entire manuscript, and George Ashton, Andrew Assur, Jerry Brown, Kevin Carey, Lawrence Gatto, Timothy Kemmis, Jay Lockhart, Thomas Munsey, Roger Saucier, Art Walz, and Robert Whalin for their helpful comments, discussions, or reviews of major sections of the manuscript. Lawrence Gatto provided the photographs used for Figures 27, 31 and 35, Timothy Kemmis those for Figures 64, 65 and 74, and James Ebersole those for Figures 67 and 68.

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EXECUTIVE SUMMARY

Literature applicable to the environmental problem of shore erosion in northern reservoirs was analyzed. Because of a lack of erosion studies in reservoirs, pertinent studies from literature on lacustrine, coastal and riverine erosion were reviewed and applied, when appropriate, as analogies to the reservoir erosion problem. This review revealed that few studies have quantitatively analyzed 1) the processes of erosion and their relative importance, especially those unique to northern regions, 2) the influence of the geotechnical properties of bank materials on such erosional processes, and 3) the parameters that can be used to predict the erodibility of a given reach of shore. Only a very limited number of studies have actually analyzed erosion and recession of reservoir shores.

A clear understanding of the causes of erosion is required to effectively minimize shoreline recession and the related impacts it causes. Erosion control is especially important because shore erosion affects the environment through the direct modification of shore and nearshore zones, including loss of land, habitat and structures, and through the introduction of sediment into the water column with ensuing adverse impacts on water quality, biological activity and sedimentation rates.

Reservoirs differ from other water bodies in several aspects related to their artificial creation (superimposing the water pool upon existing geology and slope morphology) and their daily operation. These factors can affect the location, rate and timing of erosion, as well as the intensity and relative importance of certain erosional processes. Much of the continuing erosion and shoreline recession that take place are the result of shore zones responding to, and attaining equilibrium with, the new lacustrine conditions.

The shore zone consists of two main parts - the beach and the bluff. Within these parts, processes differ in their importance, with subaerial (surficial) processes important in causing failure and degradation of bluffs, and subaqueous (underwater) and wave-related processes important in beach erosion. If the water level rises enough, subaqueous processes will act directly on the bluff and, hence, create a transitional zone that may be eroded by both subaerial and subaqueous processes. Nearshore currents generated mainly by wave activity transport the erosional products from each zone into offshore areas of the impoundment.

No single process or factor uniquely accounts for shore erosion or shoreline recession. Mass movements (e.g. rotational and translational slides, debris flows, lateral spreads), freeze-thaw and other cryogenic processes, water level fluctuations, overland flow by sheet flow, rilling and gullying, ground water processes, burrowing by animals, and the action of waves, currents and ice during storms or periods of high water are commonly processes modifying bluffs, but they may vary in importance within a reservoir.

Beach erosion results mainly from wind waves, longshore and rip currents and water level fluctuations (including those generated by storms), ice cover movement, and subaqueous mass movements. Soil moisture and

ground water conditions can significantly alter the effectiveness of erosion by tractive forces and determine shore zone stability. Processes vary in importance seasonally in northern areas, with some evidence that maximum periods of beach and bluff erosion commonly occur during late winter or early spring, and during periods of repetitive freezing and thawing.

Geotechnical properties of beach and bluff materials can determine the effectiveness of a particular erosional process. Reservoir siting in northern regions typically leads to shores composed of alluvium, and of glacial deposits, about which little is known concerning erodibility. The origins of these deposits and the resulting stratigraphic sequences may provide a systematic approach to analyzing their response to erosional processes. Vegetation in general inhibits bluff and beach erosion and stabilizes slopes. Modifications by man may retard shoreline recession at one location while enhancing it elsewhere. A condition of seasonal or perennial frozen ground affects erodibility, but current research presents contradictory evidence on the importance and nature of these effects. Ice cover may act as protection, mainly by limiting wave action, or it may bulldoze onshore sediments, and entrain and transport beach and bluff material into foreshore and offshore areas. A snow cover generally protects beaches and bluffs, but spring runoff from snowmelt may result in significant erosion.

Current understanding of beach and bluff erosion and shoreline and bluffline recession in impoundments is severely limited. Field research is needed to quantitatively analyze the processes of erosion and their relative importance, define the factors determining the nature, rate and timing of erosion, and identify factors that may naturally limit erosion in reservoirs. Erosion is clearly an environmental problem requiring interdisciplinary research in order to understand and predict its occurrence, and thus to develop appropriate techniques to minimize shoreline recession and the environmental impacts of erosion.

EROSION OF NORTHERN RESERVOIR SHORES
Analysis and Application of Pertinent Literature

Daniel E. Lawson

INTRODUCTION

Shore erosion is a problem that can directly and indirectly cause adverse environmental impacts. Physical, chemical and biological systems may be modified in areas undergoing erosion. As a result, ecological, aesthetic and cultural resources may also be adversely affected. Economically, the destruction of lake and reservoir shores and streambanks is a worldwide problem that directly causes the loss of billions of dollars in land and structures annually. The economic impacts of the indirect effects of erosion are at present incalculable, but may far exceed those of the direct effects. Costs just to arrest the erosion of streambanks in the United States were estimated in 1981 to be over $1 billion annually (COE 1981).

Because of the direct and indirect environmental impacts, erosion must often be reduced or eliminated along certain lake, river and reservoir shores. But without a clear understanding of the causes of erosion, shoreline recession cannot be effectively and economically minimized. Because of its complexity, any analysis of the physical aspects of erosion and any solutions to shoreline recession require a multi-disciplinary approach that considers geological, geotechnical, soil and engineering factors (Edil and Vallejo 1980), as well as considering their environmental impacts (Henderson and Shields 1984). Unfortunately, the study of shore zone sedimentary processes is one of the most difficult to undertake and advancement in knowledge has been slow.

In this monograph, I present the results of an in-depth analysis of literature on erosion processes, including their mechanics, and the factors affecting the erodibility of unconsolidated materials in the shore zone of northern reservoirs. Within this context, characteristics of the reservoir environment are described. The environmental impacts are also discussed in order to emphasize the importance of erosion as both a direct and indirect environmental problem. A primary intent of this monograph is therefore to assess the current state of knowledge on the causes of erosion in reservoirs, including identifying inadequacies in understanding this problem and thus areas in need of further research.

The depth of the review and breadth of this monograph were undertaken because a single source that discusses the concepts of erosion within the reservoir environment is not available. Each section is meant to be somewhat inclusive so that the reader interested in a specific process or factor need read only that section and not the entire monograph. The monograph should provide sufficient background information for a preliminary assessment of erosion problems in reservoirs.

Research on the causes and factors affecting the erosion of reservoir shores is very limited. For this reason, I have included an analysis of research results on lake, coast and river erosion for situations that are analogous to those of the shore environment of northern reservoirs. Regardless of the literature source, very few studies have quantitatively assessed the processes of erosion, factors affecting their occurrence and intensity, or their relative importance, the geotechnical properties of beach or bluff materials in relation to erosional processes, or the techniques for predicting the erodibility of northern reservoir shores.

EROSION -- IMPACTS ON THE ENVIRONMENT

Lake shores and streambanks are normally subjected to erosion by a variety of natural processes. Man's activities within watersheds and in channel systems can significantly affect these processes and modify their intensity and frequency (e.g. Vanoni 1975, Chap. 1). Impoundments formed behind dams inundate slopes of materials that were previously affected only by surficial conditions, thereby subjecting them to new lacustrine beach and nearshore processes. This alteration of the natural characteristics of the new shore zone of a reservoir most often accelerates the erosion rate. Accelerated erosion and thus accelerated sedimentation elsewhere can produce numerous environmental problems that can have significant economic ramifications and may subsequently lead to legal conflicts (e.g. Vanoni 1975, Chap. 6,7). Soil erosion, of which shore or bank erosion is a primary component, is clearly one of the world's most serious environmental problems (Toy 1982).

Adverse environmental impacts of erosion in impoundments result from 1) the direct modification of the shore zone, or 2) the introduction of the eroded sediment into the water column of the reservoir pool, including its eventual transport downstream from the impoundment.

The modification of the shore zone by erosion is the most visible of its impacts and clearly attracts public attention because of the reduced aesthetic appeal of the area (Bhowmik 1978). Erosion causes the loss of riparian property, including land and structures, and the damage or loss of roads, bridges and other structures. Legal conflicts can arise, for example, when erosion of federally owned property extends beyond it into privately owned land, and when damage results from the addition of the eroded sediment into water or from the deposition of this sediment (Vanoni 1975, Chap. 7).

In addition, erosion can modify or eliminate existing aquatic, terrestrial, and wetland habitats. In each case, undesirable shifts in production, diversity, density and composition of the vegetation and wildlife communities may result (see, for example, discussions in Currier 1954, Barton and Winger 1973, Buckley et al. 1976, Darnell et al. 1976, Shields 1982). Certain new habitats can also be created, however (Beckett 1978).

The removal of vegetation adjacent to a lake or impoundment may modify ground water and overland flow, thereby decreasing the potential filtering action and improvement of water quality (including bacterial conditions) that can occur before the water is discharged into the impoundment (Nikolayenko 1974).

Loss of shore vegetation results in further acceleration of erosion. Vegetation reduces sheet wash and rill formation on bluffs (e.g. Hunt et al. 1976), while marsh plants and other vegetation in shallow nearshore areas can effectively dampen waves and currents and limit their intensity (Pincus 1962, 1964).

Hydrologic conditions are also altered, including modification of the water table, drainage system, and location and intensity of overbank flooding. Such changes in hydrology can significantly modify terrestrial and wetland riparian habitats (Austin et al. 1979, Barclay 1980). Fluctuating water levels common to reservoirs will result in a zone of temporally variable hydrologic and biologic conditions and may affect the use of ground water for water supply by urban developments located on their shores.

The introduction of sediment into the adjacent water column has significant ramifications, perhaps the most important of which are the sediment's effects upon water quality and biological activity. Sediment has been assessed as the major water pollutant by weight and volume, as well as the primary carrier of undesirable water quality constituents such as pesticides, nutrients and pathogens (U.S. Senate 1960). Sediments affect water quality, both as a major pollutant and as a catalyzing, transporting and storage agent that contributes to other forms of pollution (Livesay 1970, Ackerman et al. 1973, Steele and Stefan 1978, Kennedy et al. 1981), with the complex hydrodynamics of reservoirs playing a major role in determining water quality (Keeley et al. 1978, Thornton et al. 1981a). Taste, odor, temperature, abrasiveness and turbidity of water are affected by suspended sediment; hence, its simple presence in water supply reservoirs can result in the need for increased water treatment (e.g. Symons 1969, Oschwald 1972, Vanoni 1975, Chap. 1, Baxter 1977).

Suspended sediments affect aquatic biota through their lowering of light penetration (turbidity) and the resultant changes in photosynthesis and primary biological production and species composition (Stall 1972, Oschwald 1972, Hecky et al. 1974, Geen 1974, Barko 1981). Fish may suffer from suspended sediment since it can obscure vision, limit feeding and reproduction, and abrade gills (Ellis 1936, Oshwald 1972). Suspended material affects the thermal structure of reservoirs by affecting its light attenuation and albedo (Dhamotharan and Stefan 1981). Deposition, including flocculation, can alter bottom habitats and thus biological activity at the water/sediment interface (Livesay 1970, Avakyn 1975, Cooper and Bacon 1981) and disrupt the food chain of the reservoir ecosystem (Baxter and Glaude 1980). Eutrophic conditions may thus result from the suspended and deposited sediments (e.g. Sefton and Meyer 1981).

The capacity of sediments to assimilate various nutrients, pesticides and heavy metals determines their importance in introducing contaminants into impoundments, with fine-grained cohesive sediments having the largest assimilative capacity (Ariathurai and Krone 1976, Kennedy 1963, Oschwald 1972, Nordin 1975, Green et al. 1978, Karickhoff and Brown 1978). Eutrophication may result when these nutrients are transported and deposited within bottom sediments of the impoundment. The gradual release of nutrients with time can lead to accelerated biological activity, causing overproduction and death of aquatic organisms, and further accumulation of these nutrients in the bottom sediments. This cycle can then start anew,

although the continuing deposition of erosional material may subsequently further accelerate, or depress, the nutrient exchange level (Howeler 1972, Baxter 1977, Hembree et al. 1971).

Pesticides such as DDT, aldrin, dieldrin, chlordane and others in agricultural areas are a particularly important source of nutrients that may cause eutrophication (Weber 1972, Holdgate and White 1977). Impoundments serve as sediment traps, capturing up to 90% of the incoming sediment (Brune 1953), and thus can be sinks for organochlorine and other residues adsorbed onto sediment particles (Pionke and Chester 1973, Ricci et al. 1983). Soil particles are estimated to transport over 90% of the organic nitrogen and phosphorus originating from erosion of agricultural uplands in an adsorbed state (Illinois Environmental Protection Agency 1978). Once released by erosion into the water column, they may be deposited in the bottom sediments (e.g. Klaasen and Kadoum 1975, Kent and Johnson 1980, Schnoor 1981, Ricci et al. 1983, Leung 1979). These bottom sediments of reservoirs can in general entomb chemical, radioactive or organic pollutants and become a long-term source of toxins (Livesay 1970, Oschwald 1972, Rea et al. 1981, Kennedy et al. 1980, Ricci et al. 1983). The ultimate fate of pesticides and other toxins deposited in impoundments remains, however, poorly understood (Leung 1979).

The sediment released by erosion decreases storage capacity and ultimately can decrease the use of impoundments for flood control, water supply, power, irrigation, recreation or other multiple purposes (e.g. Hagan and Roberts 1972, Vanoni 1975, Chap. 6, Hodgins et al. 1977, Austin et al. 1979, Kennedy et al. 1981, Sefton and Meyer 1981). The life of a reservoir depends upon the rate at which sediment is deposited within it; thus, the continued operation of a reservoir facility may depend upon maintaining the pool size through the costly removal of this sediment. An indication of the potential problem of eroding shores as a sediment source is that of Rea et al. (1981) who concluded that over twice the amount of sediment deposited within the Great Lakes (about 64% of the total) is derived from shore erosion than is derived from atmospheric sources and from the bedload and suspended load of tributary rivers. Similarly, Evans and Schnepper (1977) and Leedy (1979) estimated that 50% or more of the suspended sediment in Illinois streams is derived from channel erosion. Avakyn (1975) estimated that millions of metric tons of sediment are introduced annually into Soviet reservoirs from bank erosion. Studies within newly formed impoundments have documented similar high inputs of sediment from shore erosion (e.g. Van Everdingen 1969, Newbury et al. 1978).

The location of deposition of the eroded sediment can cause acute problems as well, particularly within the smaller and shallower reservoirs. Shoaling of inlets and coves by redeposited material can destroy their function as boat harbors and launches or as recreational areas.

Additional effects of erosional losses may include an increase in the surface area of the impoundment as shorelines recede, thus increasing the potential loss of water by evaporation and contamination (Baxter and Glaude 1980).

THE RESERVOIR ENVIRONMENT

The reservoir environment differs significantly from that of lakes and rivers (e.g. Lowe-McConnell 1966, Ackerman et al. 1973, Baxter 1977, Thornton et al. 1981b, Stefan 1981), although the limnological and physical processes that affect lakes also affect reservoirs. Most differences in the environment owe their origin to the artificial creation of the impoundment and to the day-to-day operations of the reservoir, operations that are determined by its principal uses. The reservoir, however, still remains a part of a system -- the drainage basin of which it is an integral part. Events occurring within the reservoir are affected by events upstream of its location, and similarly, events within the reservoir have effects downstream. Thus in the context of erosion, numerous factors may interact to erode a particular reach of the shore, yet they are also a function of basin-wide factors that are external to the immediate shore environment (as an example of this concept, see Walling 1983).

The classic work of Hutchinson (1957) is an extensive treatment of the origins, physical characteristics, hydrodynamics, thermal structure and other limnological aspects of the lake environment and a source of more detailed information than can be presented in this monograph. Additionally, more recent research on the limnology and hydrodynamics of lakes is discussed in Csanady (1975), Mortimer (1974), Graf and Mortimer (1979), Imberger and Hamblin (1982) and Lerman (1978), among others. Fox et al. (1979) and Bogardi et al. (1981) have reviewed literature on modeling of northern lake processes.

In most reservoirs, the water level is manipulated according to the purpose for which it was built, but it is governed by local climatic and hydrologic constraints that also normally cause changes in water levels in natural lakes. Water levels in lakes may undergo seasonal, short-period and long-period fluctuations. Seasonal variations result mainly from varying amounts of precipitation, evaporation, ground water flow (in and out of the basin) and surface runoff. Water levels are generally higher in summer and lower in winter. A rising level during early spring can result from snowmelt runoff and heavier-than-average rainfall.

Short-period fluctuations in water level (temporary displacements of the water surface) are caused by wind and barometric pressure differences, and the resulting seiche (Einarsson and Lowe 1968). Wind stresses due to storms on large lakes or reservoirs can drive water onto the shore zone at a rate faster than subsurface currents can move this incoming water away from it, thus causing a stationary oscillation of the surface ranging from a few millimeters to several meters (e.g. Hamblin 1976, Saville et al. 1962). Water levels on opposite sides of the lake therefore differ; in Lake Erie, this may amount to a change of 4 m from the upwind western edge to the downwind eastern end. Obviously, the size of reservoirs precludes differences of this magnitude. Barometric pressure changes due to the closely spaced passage of high and low pressure storm systems can cause a similar tilting of the water surface. Larger seiches often result from a heavy downdraft associated with squall line thunderstorms. In all cases, diminishing wind stress or pressure allows a surge of the water toward the lower side of the lake. This in turn produces a surge or flow of water in the opposite direction and continued repetition of this surging generates

a seiche. Its magnitude depends upon the original force and amplitude of the standing wave that is formed (Hunt 1959).

Long-period fluctuations in water level (several years or more) result mainly from long-term changes in climatic parameters that affect the annual amount of precipitation. Drought conditions can cause severe changes in the reservoir environment, including modified thermal stratification, nutrient loading and increased turbidity, because of increased mixing and turbid conditions caused by wind stress on the shallower water (Rettig 1981, Cassidy et al. 1981, LaBounty and Sartoris 1981).

Superimposed upon these natural fluctuations in the water level of reservoirs are those resulting from reservoir operations. Most reservoirs are filled at times of high discharge, and are either drawn down gradually in accordance with the reservoir's primary purpose before the next high discharge, or maintained at a constant level, with a rapid drawdown just prior to the next high discharge (Baxter 1977). Certain reservoirs built solely for flood control are filled only during flood events and remain empty for the remainder of the time. Reservoirs having several primary uses may have far more complex patterns of water level fluctuation resulting from operations.

Reservoir morphology

Because dams are generally located within river valleys, the shoreline configuration, bottom topography and morphology of the reservoir often reflect the morphology of the river and tributary river valleys before inundation. High and narrow valleys result in long, narrow and often sinuous reservoir shapes, whereas in regions with much less relief, the impoundments may spread and form a pool over a wide area. Damming the outlet of a lake results in a much less pronounced difference in shoreline morphology.

Reservoirs therefore typically differ from natural lakes in having greater shoreline development (ratio of shoreline length to circumference of a circle of the same area as the reservoir pool) and skewed longitudinal profile (Baxter 1977). River reservoirs are typically deepest just upstream of the dam whereas natural lakes are usually deepest near their centers. This bathymetry affects hydrological processes, as does the withdrawal of water from this deep area through dam outlets that may be located at various depths below the water surface along the vertical axis of the dam. Inflowing currents and circulation patterns may also be affected by the former thalweg of the river (e.g. Kennedy et al. 1981).

Also, steep valley walls before inundation become steeply floored sections of the impoundment. Nearshore zones may be characterized by rapidly increasing depth over rather short distances from the shoreline, which can promote offshore movement of eroded materials (Savkin 1975) and subaqueous failures after flooding (Tarverdiyev 1972).

Once flooded, sediments composing the bed and banks of the reservoir are relicts of conditions prior to reservoir and dam construction (Kachugin 1966, 1970). The sediments beneath northern reservoirs are varied but often consist of alluvial deposits, glacial and associated glaciofluvial, glaciolacustrine and periglacial deposits, and colluvium and other slope materials. Their geotechnical characteristics, which are derived from

their origins and from post-depositional processes such as weathering, will have a strong effect on their immediate and long-term responses to the new lacustrine conditions superimposed upon them. The general anisotropic nature and hydraulic properties of the material can exert considerable control over these responses to flooding (Wahlstrom 1974). Although stable previously, that stability can be significantly reduced during reservoir filling, and subaqueous and subaerial failures may be triggered. Glacial deposits are generally unique to northern regions; their potential interaction with erosional processes will be described later in this monograph. Fisk's (1952) description of the interaction of alluvial deposits with erosional and depositional processes in rivers illustrates the concept of bed and bank adjustments in relation to sedimentary origins of shore zone material.

Because the newly created shore zone is not in equilibrium with the lacustrine environment, it represents an unstable configuration. In general, physical changes to the shore zone will be the product of the energy of the eroding forces, the geometry of the reservoir, surrounding terrain and shore zone, and the resistance of the bank materials to the different erosive forces (Krumbein 1950, Baxter 1977). The time necessary to reach an equilibrium will vary from reservoir to reservoir as well as from reach to reach within a single impoundment. Because of a lack of understanding of the erosional processes and their interrelationships with one another and with the morphology of the shore and nearshore environment, it is not yet possible to accurately calculate the number of years required for equilibrium to occur, although estimates have been attempted for specific locations where only one or two erosional processes and factors were considered important (e.g. Black 1980).

Several authors have suggested that the development of eroding slopes and shore zones can be considered an evolutionary process (Kondratjev 1966, Kachugin 1966, Brunsden and Kesel 1973, Quigley et al. 1977, Newbury et al. 1978, Hands 1980). This process in reservoirs conceptually involves the volumetric displacement of sediment at or above the water line to some point below it in order to establish a stable, equilibrium profile. These adjustments are made in accordance with the hydrodynamics of the reservoir, as they are within river systems (Fisk 1952). This concept is essentially that proposed by Bruun (1962) for the erosion of coastlines due to the rising water levels of oceans; it will be discussed in more detail in a subsequent section of this monograph.

Variation in the size, shape and orientation of impoundments can directly or indirectly affect the relative importance and intensity of reservoir processes (e.g. Stefan and Demetracopoulos 1981). For example, the shores of an impoundment that fills a valley aligned generally along the principal wind direction will have a long fetch, and shore zones will be more intensively attacked by wind-generated waves than will shores oriented perpendicular to the principal wind direction (e.g. Savkin 1975). Similarly, a reservoir of complex shoreline configuration may have sections of shore within hundreds of meters of one another that are differently affected by wind-generated waves, with promontories more susceptible to attack than bays. In this regard, Pincus (1962) discussed the variability in erosional and depositional processes that are active within six distinct shore environments of the Great Lakes.

Some authors have attempted classifying reservoirs on the basis of their predominant morphological characteristics (e.g. Edel'shteyn 1977). Lara (1962), for example, suggested that the shape of the reservoir, of which he classified four types, could be used to predict the general distribution of sediment within it. Other characteristics, such as regional topography, landscape, operating conditions, water level fluctuations, reservoir head, usable storage and water exchange rate, are included in Podlipskiy and Shirokov's (1976) classification. They considered these variables extremely important in determining the hydrologic regime of large reservoirs in Siberia.

Physical limnology

River inflow and water outflow at the dam often predominate in determining the retention time of water and circulation within reservoirs, in contrast to natural lakes in which wind-generated currents and thermal circulation determine retention time and current patterns (Baxter 1977, Thornton et al. 1981a,b). The water budget of reservoirs may include 1) surface waters, usually streams, 2) ground water inflow, 3) ground water outflow into a permeable bed under the proper head, 4) evapotranspiration, and 5) bank storage, which will vary with changes in the water table that result from fluctuations of the impoundment's water level (Virkulina 1977).

Reservoirs can significantly alter the regional hydrogeologic regime and result in a long-term, permanent change. During the initial filling, a transient flow system develops with influx of water from the reservoir into shore sediments and the regional ground water flow system (Simons and Rorabaugh 1971), in a manner similar to the mechanism of bank storage in rivers during flood stages (Todd 1955, Cooper and Rorabaugh 1963). The result is that water tables are generally higher, the typical regional discharge pattern of ground water into the valley is reduced in rate, and the hydraulic heads in adjacent aquifers are increased, each in accordance with the final water level of the reservoir (Freeze and Cherry 1979). The magnitude and rate of water level fluctuations, once a reservoir is filled, determine the magnitude of associated fluctuations and flow of ground water throughout the shore zone and adjacent land.

Nearshore currents and thermal stratification are important factors affecting erosion, deposition and the development of offshore and foreshore bottom topography. In particular, sedimentation in reservoirs depends upon the quantity, time, distribution, and particle size of the sediment source, and the size, shape and operation of the reservoir (Wiebe and Drennan 1973). The importance of advective and unidirectional transport in reservoirs results in pronounced but variable physical, chemical and biological gradients (Thornton et al. 1981a). Suspended sediments introduced from the shore zone are distributed and deposited within the impoundment in accordance with these gradients, as well as by currents and wind waves (Johnson 1981).

Because of distinct temperature differences between summer and winter and the water density at these different temperatures, lakes and artificial impoundments of northern temperate regions are often characterized by a three-layer thermal stratification (Hutchinson 1957, Kittrell 1965). An upper region, known as the epilimnion, consists of relatively warmer, free-

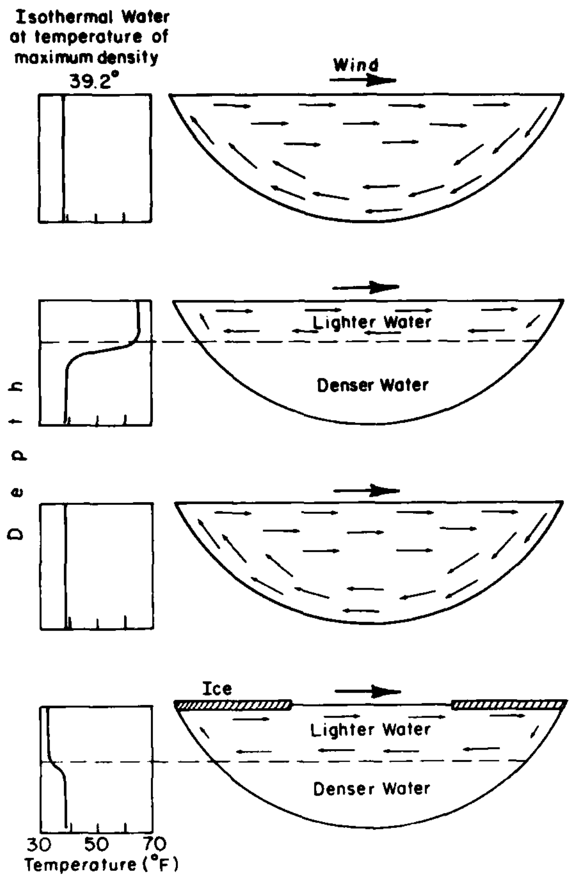


Figure 1. Idealized thermal stratification and seasonally induced circulation patterns of northern lakes and reservoirs (after Hough 1958).

ly circulating water with a small temperature gradient. The lower, colder and much denser region, called the hypolimnion, consists of poorly circulating water at temperatures of near 4°C or below throughout the entire year. The region separating the two layers, the metalimnion, is defined as the zone of maximum temperature gradient and marks the limit of vertical mixing associated with surface stresses. In some cases, the metalimnion consists of two layers: an upper zone of intense thermal gradient, the thermocline, and a lower zone, the clinolimnion, which also has a steep thermal gradient but has other properties similar to those of the hypolimnion (Hutchinson 1957).

Physical modifications to this three-layer structure result mainly from wind- and wave-induced currents and internal waves (e.g. Ragotzkie 1978, Csanady 1978, Imberger and Hamblin 1982). Wind stress on the water surface causes turbulent mixing of the epilimnion and, in shallow impoundments, can control both thermal and circulation patterns (e.g. Stefan and Demetracopoulos 1981). Within reservoirs, currents developed by discharge at the dam, the temperature of inflowing water, and prolonged periods of lowered water levels (e.g. Rettig 1981), may also alter the temperature and water circulation patterns (e.g. Wunderlich and Elder 1969, Ashton 1980, Imberger 1980).

In most northern lakes and many northern reservoirs, there are periods during which the entire body of lake water undergoes circulation because of seasonally induced water temperature variations (Fig. 1). Al-

though such overturns usually occur seasonally, local climate and reservoir morphology can result in circulation at other times of the year.

Full circulation usually begins in spring after the ice cover is lost. After overturn, mixing of the lake waters briefly results in an isothermal condition for the entire water body at 4°C, the temperature at which water has its maximum density. Thermal stratification develops as warmer air temperatures, solar radiation, inflowing stream water and surface runoff produce warmer surface waters that become sufficiently buoyant and deep to resist the complete vertical mixing of the impoundment by the wind. The colder waters below this relatively shallow surface layer resist mixing and a vertical temperature gradient between the surface and deep water develops, forming the hydrostatically stable thermocline.

Once stratification has developed, the warmer epilimnion is generally kept in motion by the wind. Surface waters flow as a result of the shear stresses exerted by the wind, with a return flow set up within the lower epilimnion because of the deflection of the surface water mass at the shore and the resistance of the underlying colder strata to mixing. Continued heating of the surface waters during summer drives the metalimnion deeper, greatly inhibiting turbulent mixing or currents within the hypolimnion. Warm periods with extensive mixing that alternate with cool periods during summer storms may produce a step-like pattern in the temperature gradient and two or more thermoclines. Circulation within the epilimnion maintains a rather uniform temperature commensurate with the seasonal mean temperature whereas the hypolimnion remains around 4°C during the summer.

In autumn, there is a net heat loss and the temperature of the epilimnion gradually drops. Thickening of this stratum takes place, while the cooler and denser surface water descends convectively throughout the epilimnion. The density and temperature of the upper layer approaches that of the hypolimnion; eventually, wind-induced circulation encompasses the entire lake and an isothermal condition is again attained. As temperatures continue to drop, the relatively small differences in density of waters at temperatures below 4°C, and eventually nearing 0°C, allow relatively easy mixing of the entire water body by the wind. An ice cover forms when water temperatures are at or below 0°C and calm conditions are accompanied by strong outgoing radiation and evaporative and sensible heat loss to the atmosphere (Michel 1971). The larger the lake, the longer must be this period of cold and calm before the ice cover's integrity is sufficient to resist wind stresses and thus break-up. Ice formation can occur rather rapidly and, in a small water body, may cover it entirely in a few hours (Ragotzkie 1978). In contrast, in very large lakes or those with long fetches, an intact ice cover may not form and open water areas may remain through most or all of the winter, thus permitting other types of ice, such as frazil ice, frazil slush and pack ice, to develop and also accumulate along the shoreline (e.g. Michel 1971). Which forms of ice develop depends upon such factors as the wind speed and direction, air temperature, incoming and outgoing radiation, and water temperature (e.g. Michel 1971, Ashton 1980). Ashton (1980) briefly discusses a method based upon meteorological variables that appears reasonably successful at predicting the ice types that will form along the shorelines of large lakes.

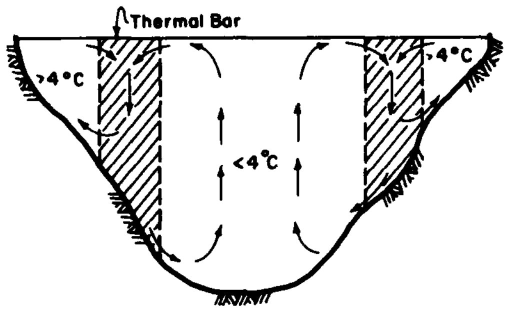


Figure 2. Idealized sketch of thermal bar and associated circulation pattern as it develops in the Great Lakes in spring (after Rodgers 1965).

A complete ice cover does not alter the thermal stratification of the lake but can restrict circulation (e.g. Michel 1971). The upper layers of water in contact with the ice remain at 0°C, while thermal diffusion at the sediment/water interface maintains temperatures at the bed above those of the epilimnion. Solute diffusion may further increase the density of the hypolimnion, while partial melting of the ice layer injects rather clean water at 0°C into the uppermost strata of the lake. Both processes tend to maintain or possibly increase density differences between the epilimnion and hypolimnion. This warming to near 4°C may not take place in reservoirs with sufficient throughflow during winter (Ashton 1980); natural river inflows are typically near 0°C and, depending upon the inflow discharge and the rate and location of withdrawal at the dam, can maintain water temperatures near 0°C.

Seasonal warming in spring leads to melting, candling and breakup of the ice cover. Progressive melting with seasonal warming often first starts along the shore, where ice is thinner and abutting the warmer ground (e.g. Williams 1966). This may develop open water and leave the ice cover floating, free of the shore. As the ice surface continues to melt, this meltwater flows into cracks and through the ice itself by preferential melting of ice grain boundaries (Michel 1971, Gow and Langston 1975, 1976). The structure and integrity of the ice cover are gradually lost. At this point, wind stresses can easily break up the ice cover and induce circulation of water at depth. As the input of solar radiation and air temperature increases, convective mixing begins and the entire lake warms to near 4°C. The thermal stratification cycle then begins anew.

Rodgers (1965, 1966) found that temperature gradients and the cycle of thermal stratification are not necessarily this simple. In the Great Lakes, a thermal bar or vertical mixing zone develops during the final stages of the spring overturn because nearshore water warms to 4°C sooner than the central part of the lake (Fig. 2). Thus, a vertical mixing zone lies between mid-lake waters that are less than, and near shore waters that are greater than, 4°C. Surface waters converge at the bar and they sink, while upwelling takes place in mid-lake waters. Horizontal currents caused by thermal gradients are restricted to areas shoreward or lakeward of the bar. Warmer river water and surface runoff entering the lake may influence where bar formation takes place, as well as its general form and position. Because nearshore water is confined to the edge of the lake basin in early spring and moves offshore only at the rate of bar movement, river effluent is held within this nearshore zone (Noble and Anderson 1968). Very little

mixing occurs between the river plume, the bar zone waters and the central open lake water.

As spring warming progresses, the thermal bar moves offshore toward the center of the lake in a manner analogous to a contracting cylinder. Bar movement appears to depend on the depth, bottom topography and temperature gradients on the cold side of the bar as well as the rate of nearshore temperature increase (Rodgers and Sato 1970). A thermocline develops in nearshore waters, while water temperatures lakeward of the bar remain homogeneous; development of stratification throughout the lake thus depends entirely upon the lakeward movement of the bar (Rodgers and Sato 1970).

Whether thermal bars characterize deeper and larger reservoirs in spring is unknown. Their presence would have important effects on the movement of suspended sediment by nearshore currents and river discharge. Thermal bars might occur in reservoirs where inflowing river water maintains temperatures near 0°C during winter and where a zone of shallow water lies near the shoreline.

Various hydrodynamic models of three-layer thermal stratification and circulation within reservoirs have been proposed following the basic principles described above (e.g. Stefan and Ford 1975, Johnson 1981, Harper and Waldrop 1981, Brown 1981, French et al. 1981), but they have not yet been adequately tested on a local or regional basis.

Thermal variations within impoundments are important because they can affect the transport of sediment and determine the timing and locations of suspended sediment deposition (Dhamotharan and Stefan 1981). These variations are thus directly related to water quality (Dhamotharan and Stefan 1981). Viscosity variations with changes in water temperature in a reservoir can affect the settling rate of particles, and thus the turbidity of the water column, and may affect the ability of currents to entrain and transport bed and bank material.

Most researchers agree that a decrease in water temperature increases the rate of suspended sediment transport by unidirectional currents if other factors such as discharge and sediment characteristics remain constant (e.g. Straub et al. 1958, Ali 1961, Toffaleti 1968, Colby and Scott 1965, Mellema 1970, Taylor and Vanoni 1972a, b). This increase results mainly from the increase in viscosity of water as the temperature decreases -- viscosity is approximately twice as great at 0°C as at 25°C. This in turn causes a decrease in the fall velocity of particles suspended in the water (e.g. Straub et al. 1958, Colby and Scott 1965, Mellema 1970). A similar effect results from high sediment concentrations, which also increase the viscosity of the sediment/water mixture and can decrease the fall velocity of particles (Colby and Scott 1965). In a reasonably controlled field study of a portion of the Missouri River, Shen et al. (1978) concluded that, as temperatures decreased, the respective increase in fluid viscosity decreased the rate and magnitude of suspended particles' fall velocity, thereby increasing the rate of suspended sediment transport. The effect of water temperature on bedload transport is less well-defined but may be similar (e.g. Shen et al. 1978, Franco 1968, Taylor and Vanoni 1972a, b, Colby and Scott 1965). Thus, the erosiveness of nearshore currents and waves (their ability to entrain and transport sediment) may vary seasonally with changes in nearshore water temperature.

The implications for reservoir sedimentation are clear - lower water temperatures will increase the length of time that sediments from eroding shores remain in suspension and possibly their distance of transport from the eroding bank. Thermal zones will likewise affect the location and duration of suspended sediment transport. Seasonal variations in sediment transport should therefore be expected in artificial impoundments affected by seasonal temperature variations and thermal stratification.

Even in impoundments without well-defined thermal strata, a density stratification may result from a linear temperature gradient that extends from the water surface to the lake bottom and affect reservoir sedimentation patterns (Kittrell 1965). Superimposed upon this vertical temperature gradient may be a longitudinal gradient that results from a current structure established by inflowing stream waters and by discharge at the dam (Baxter 1977, Markofsky 1979, Thornton et al. 1981a).

The effect of thermal stratification on currents and sediment dispersal is clearly demonstrated in the case of density flows, where inflowing water may enter an impoundment as an underflow, interflow or overflow in accordance with the density differences of the inflowing water (e.g. Wiebe 1939, Anderson and Pritchard 1951, Slotta 1973, Wunderlich and Elder 1970, 1973, Ford and Johnson 1981, 1983, Imberger and Hamblin 1982). Large density differences due to sediment concentration can produce turbidity currents that result in resedimentation of large quantities of materials (e.g. Gould 1951, 1960). Seasonal changes in the position of temperature strata, particularly during the seasonal warming beneath an ice cover in spring, may initiate density flows that move from nearshore regions into deeper offshore waters as the nearshore waters warm to near 4°C (Hutchinson 1957). Conversely, a reservoir pool that is warmer than inflowing or outflowing water can result in interflows in the hypolimnion and rapid movement of this water through the impoundment. Lateral mixing of density flows within the impoundment is apparently a function of pool morphology (Ford and Johnson 1981). Smith (1978), Hamblin and Carmack (1978), and Pickrill and Irwin (1983) have established the importance of cold inflowing water with relatively low fine-grained sediment concentrations in generating overflows or interflows in glacially fed, thermally stratified lakes. Smith (1968) found that the movement of overflows and interflows within these lakes responded to wind-driven currents and Coriolis effects.

Rip, longshore and other nearshore currents may be influenced by thermal density differences as well, particularly in areas of complex bottom topography where deeper and colder waters lie close to the shoreline. Currents within impoundments can be complex, with their location and importance varying with physical and thermal properties of the reservoir. The principal properties of a reservoir that affect currents include stream influx, thermal stratification, the height and volume of discharge at the dam, the geometry, dimensions and bottom topography of the reservoir pool, and external inertial forces such as wind (e.g. Hutchinson 1957, Wunderlich and Elder 1970, 1973, Granju et al. 1973, Slotta 1973, Savkin 1975, Csanady 1978, Kennedy et al. 1981, Johnson 1981, Imberger and Hamblin 1982). Deep water currents and the overall circulation pattern will determine the ultimate fate of reservoir sediments. Other factors that determine the actual patterns of most sediment deposition include the size and mineralogy of the sediments, dissolved mineral content of the water, and total sediment input and discharge.

Nearshore currents are important in transporting sediment from an eroding shore and thus in determining its rate of removal and subsequent deposition in other parts of the reservoir (e.g Kondratjev 1966, Van Everdingen 1969). In lakes, coarse-grained sediments tend to be confined to shallow, nearshore zones where wave- and wind-generated currents and turbulence are most active. Wave energy appears important in determining nearshore sediment distribution (Rossmann and Seibel 1977), but it may also be important in determining sedimentation patterns throughout a lake or reservoir basin (Johnson 1981). For example, fine-grained bottom sediments can be resuspended in offshore areas of shallower lakes as the result of turbulence from waves and bottom currents (Sheng and Lick 1979). Fine-grained materials are usually deposited from suspension within quiet offshore waters, or within protected coves and bays, with clay-size sediment requiring essentially motionless conditions for deposition (Sly 1978). Sediments generally decrease in grain size from the landward edge of the shore zone to the basin center. Turbidity currents, underflows or subaqueous mass movements may, however, transport coarser material into deep offshore zones.

Sediment distribution within reservoirs follows the lake pattern, with the exception that deep currents produced by river influx or by discharge at the dam may result in coarse-grained particle deposition or the formation of coarse lag deposits in deep water areas. Three zones of sedimentation have been distinguished in some reservoirs (Karaushev 1964, Wiebe and Drennan 1973). Within the zone at the upstream end of the impoundment, delta formation takes place at the mouth(s) of the inflowing river(s). Most coarse sediment transported by incoming streams is deposited in deltas because of the drop in flow velocity as the stream enters the impoundments. Typically, gravel-size sediment is deposited at the beginning of the delta and progressively finer-grained sediment is deposited basinward along the delta's long axis. Deltaic processes are often complicated by water level fluctuations, with low water levels leading to erosion of the shallow water deposits.

In an intermediate zone beyond the delta, sedimentation takes place from waves and wind- and residual river-induced currents, including density currents. Most sediment deposited here is derived from river discharge, with less from eroding reservoir shores, and consists of silt-size material.

The lowermost zone nearest the dam face apparently receives most of its sediment from the eroding banks of the impoundment (Wiebe and Drennan 1973) and is generally clay-size material deposited from suspension.

SHORE ENVIRONMENT

The shore zone is the interface between the water of the impoundment, the adjacent land, and the air. The complexity of this dynamic environment is one reason why shore erosion is difficult to analyze, and why, in general, so few recent advances in knowledge of shore zone sedimentary processes have been made. Morphologically the shore environment consists of several elements with certain erosional or depositional processes commonly active within them. Subaqueous, subaerial, or subsurface processes, or more often combinations of these, are active in causing change within the shore zone.

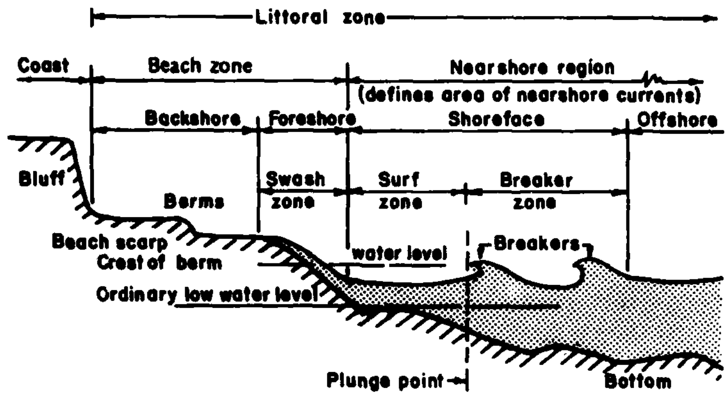


Figure 3. Schematic representation and terminology of the typical shore zone profile (after USACERC 1984).

In the discussion on reservoir erosion which follows, I will refer to elements of the shore zone profile with standard terminology developed for the coastal environment (Fig. 3). The most widely used terminology for coasts is basically that of the Beach Erosion Board (1933) with some modifications that have arisen since that time. Much of this scheme originated with the classic work of Johnson (1919). An important variation in the terminology is the concept that the beach zone includes the underwater section of the profile acted upon by breaking waves. This distinction is important because it includes the area where beach-forming processes are active; this usage is widely adopted today (e.g. Komar 1976, Davis 1978, Friedman and Sanders 1978, USACERC 1984). Various terms are defined in Appendix A.

The profiles of eroding reservoir margins often differ from those of typical coastal zone profiles in having very prominent, abruptly rising bluffs adjacent to a foreshortened beach profile (Fig. 4), with the crests of these bluffs marking the landward limit of erosion. In extreme cases, the backshore zone is absent, while the foreshore zone is severely reduced in width. At these locations, the high (or even low) water level may lie on the bluff face, at or above the bluff toe (Fig. 5). The topography of the nearshore zone is also often characterized by rapidly increasing depth within a short distance of the shoreline, unlike that of the classic beach.

The various differences in the profiles of coastal and lake shores and those of reservoirs are apparently related to the immaturity of reservoir shores. Coastal regions and lake shores are geologically much older and, although coastal or lake shores may be in a state of change, they most often appear to be in a dynamic equilibrium with the predominant sedimentological processes at their location (e.g. Fenneman 1902, Bruun 1954, Tanner 1958, Eagelson et al. 1963, King 1972, Aubrey 1979, Komar 1976). This is not true for most reservoir shore zones which appear to be out of equilibrium with these processes.

Because of this dynamic equilibrium, shore zones of lakes and coasts can respond to changes, such as climate or water level, by rapidly adjusting to a new quasi-stable configuration. Such a change is exemplified by

Figure 4. Eroding bluff (≈ 3-m height) with shortened backshore and foreshore exposed by lowered water level.

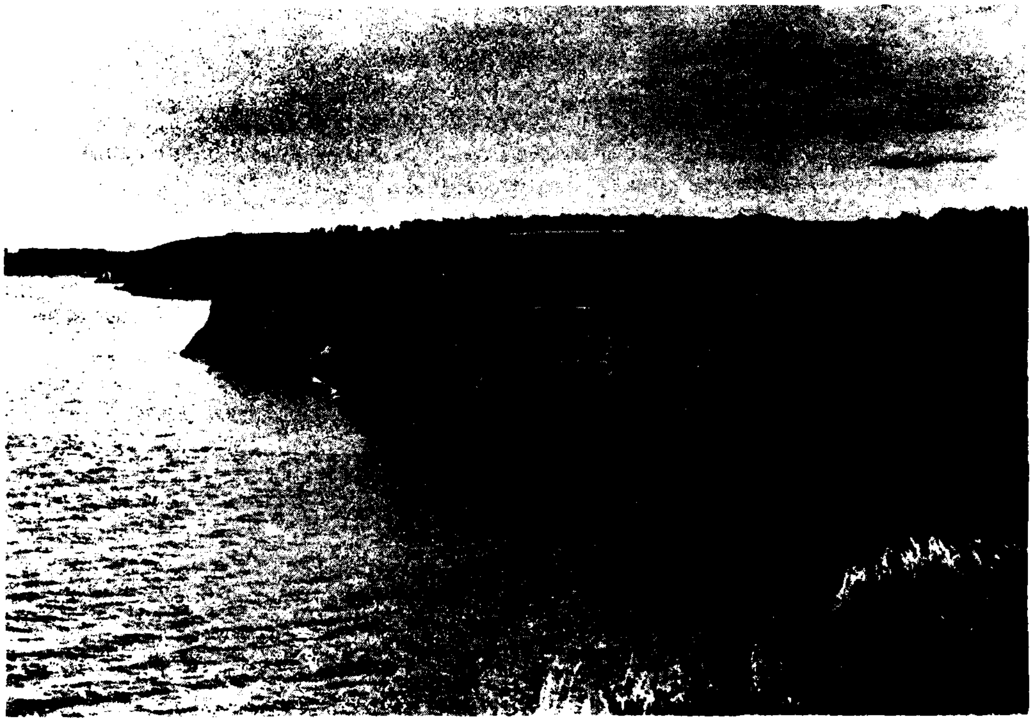


Figure 5. Deep water lying directly on base of bluff along Lake Sakakawea, an impoundment on the Missouri River, in North Dakota.

the Great Lakes where shore erosion increased markedly in response to rising water levels from 1967 to 1976 (Saylor and Hands 1970, Hands 1976, 1979, Acomb et al. 1977). Since that time, Hands (1979) found that water levels have declined and shore erosion and recession have rapidly slowed or ceased, with some beach areas actually undergoing accretion of sediment (Berg and Collinson 1976).

An understanding of the dynamics and properties of natural beaches is critical to understanding the development of the littoral zone of reservoirs. Studies of the characteristics of the shore environment of lakes and coasts have been made by thousands of researchers, and as a consequence, a large number of texts have been written on this environment (e.g. Bascom 1964, Ingle 1966, Bird 1969, King 1972, Ippen 1966, Muir Wood 1969, Hails and Carr 1975, Davis and Ethington 1976, Komar 1976, 1983, USACERC 1984). The direct application of much of this research to reservoirs is unfortunately not yet possible, but certain fundamental concepts on currents, waves and related nearshore processes are applicable and these will be discussed in subsequent sections.

EROSIONAL PROCESSES AND SHORELINE RECESSION

Before proceeding further, the use of the terms erosion and recession needs to be clarified. Erosion is distinct from recession. Erosion is a mass concept involving the net removal of a certain volume of material. Shoreline recession is a geometric concept that involves the landward displacement of the water line (Fig. 6). Shoreline change may therefore result from simply a fluctuation in water level or an actual removal of beach material. Thus, a ground or aerial survey measures recession rates and infers that erosion has taken place.

In terms of erosion, reservoir shores can be separated into two morphologic components that are acted upon by distinctly different erosional processes: the beach zone and the bluff zone (Fig. 6). Within both zones, certain processes are generally active in loosening material to be transported away from the bluff or beach faces. Because these processes may act independently of one another, bluffs may erode and recede independently of shoreline recession (or advance).

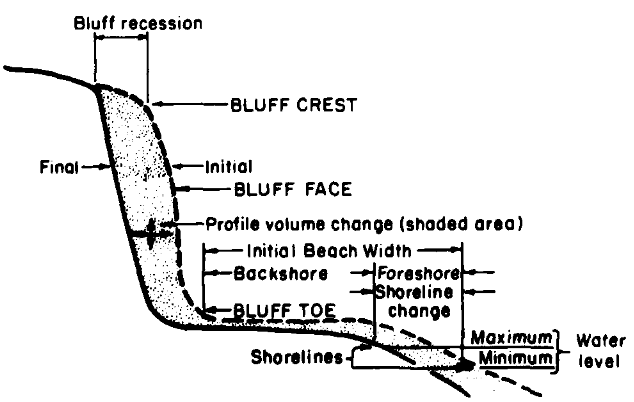


Figure 6. Concepts of erosion of bluff and beach zone sediments and recession of shorelines and bluffs (after Hands 1980). Bluff zone and beach zone erosional processes often differ and are distinguished morphologically in the text for purposes of discussion.

Reservoir operations commonly result in water level fluctuations and the usual effect is to widen or move the wave zone up or down the shore profile. This translation results in a third zone distinct from the bluff or beach zones that may be attacked by both subaqueous and subaerial processes during the course of a year. The length of time a particular process is operative within this transitional zone depends upon the fluctuations in pool height. Thus in flood control reservoirs, for example, erosion can correlate with peak periods of precipitation and flooding. Such fluctuations also modify the position of the water table, which in turn affects the potential for ground water-related erosion and shore zone instability.

As will be clear from the discussion which follows, numerous processes, as well as factors that affect their intensity or occurrence, interact to cause beach or bluff erosion and thus shoreline recession. No single process or parameter uniquely accounts for erosion or recession of a bank or bluff.

WAVES

Waves produced by stresses exerted by a wind blowing across the water surface are called wind waves. Wind waves, among all erosional processes, are most often cited (although often without substantiation) as the primary cause of erosion and recession of lake and coastal shorelines. They are the principal source of energy input to the littoral zone and vary in height from a millimeter or less to 15 to 20 m on the largest seas and lakes. Other surface waves include those created by boats and, in large water bodies, by tidal forces.

Internal waves, those occurring between water masses of differing density (e.g. in the thermocline), are basin-scale motions whose causes and properties are poorly understood (Mortimer 1971, Roberts 1975, Garrett and Munk 1979, Imberger and Hamblin 1982). Laboratory studies suggest that internal waves can break and entrain sediment in shallow water above bottom slopes or rises (Southard and Cacchione 1972). Vertical and horizontal exchange of heat and dissolved or suspended substances between thermal strata may also result from turbulent currents generated by extremely large wind stresses and seiching (Hutchinson 1957).

The relationship of basin scale motions of internal waves to nearshore erosion and deposition are basically unknown. Extreme rises in water level due to seiching could be an important "catastrophic" event that results in rapid, but short-term erosion and shoreline recession, while the piling up of water on the leeward side of a reservoir by continuous wind stresses (or perhaps barometric pressure changes associated with storms) will cause a water level rise and hence move the location of wave or current attack up or down the beach face (Savkin 1975). Because of the absence of information on internal waves, they will not be discussed further in this report.

Wind wave character and mechanics

The generation and mechanics of the orbital motion and sediment transport of natural wind waves are extremely complex and difficult to analyze.

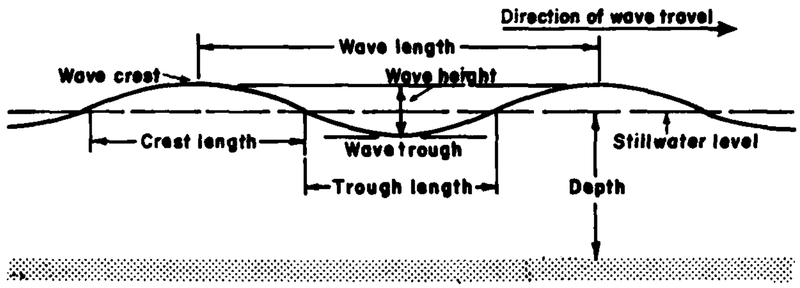


Figure 7. Definition of terms describing characteristics of oscillatory wind waves in deep water.

For this reason, theoretical treatments of ideal waves have been required to deduce their behavior. Because of natural variabilities, each wave theory assumes an idealized configuration for wave form and motion, and is limited in application to wave generation under specific boundary conditions. Numerous publications summarize this research (e.g. Wiegel 1964, Meyer 1972, Le Mehauté 1976, Kinsman 1976, Komar 1976, 1983, USACERC 1984, Muir Wood and Fleming 1981). The Shore Protection Manual (USACERC 1984) is a particularly good discussion of coastal engineering, including techniques for estimating various wave properties and wave interactions with the coastal zone. Nearshore sediment entrainment and movement by waves, current generation by wind waves, and related longshore sediment transport are, however, still only understood in mainly a qualitative sense (Horikawa 1981).

The interaction of waves with nearshore and coastal sediments and the propagation of littoral currents are particularly important for shore erosion. Waves are generally described in terms of the spectra of wavelength, wave height and wave period (Fig. 7). Statistical evaluation of these spectra define the significant wave height H_s and the significant wave period T_s, two of the most useful for describing waves (Allen 1982). H is the average height of the highest one-third of waves measured over a given period of time, and T_s is the average period of these highest waves. These variables are significant because they relate to wave energy, the ability of waves to apply force to the shore zone. Since wave energy is proportional to the square of wave height, H_s measures the modal energy of a particular series of waves (Allen 1982).

In addition, the wave climate, or the seasonal changes and character of the wave spectrum, may influence the effect of waves on coastal zone erosion. Wave climate can be determined by field measurements over long time periods, or defined theoretically by analysis of the local geography and meteorology. Numerical hindcasting procedures are used to evaluate wave climate based upon wind data and other related parameters at a site to produce a wave energy spectrum and predict significant wave height (see for example the discussion in Komar 1976, pp. 81-95). But actual field measurements are often hampered by bad weather, while the mechanics of wind wave generation are not sufficiently understood to develop fully accurate predictive models (Allen 1982).

The character and intensity of wind waves approaching the shore zone generally depend upon 1) wind speed and direction, 2) storm/wind duration, 3) the effective fetch (the length and width of the water surface over which the wind blows), and 4) water depth (e.g. Bhowmik 1976, USACERC 1984, Muir Wood and Fleming 1981). The longer and harder the wind blows, the larger the waves and the longer the period of time for decay of wave systems after the wind dies. Fetch governs the area of the water surface affected by the action of the wind and thus the time over which wind energy is converted into wave energy. It also limits the period and height of waves generated and is affected by landforms adjacent to the basin as well as basin geometry (e.g. Saville 1954, USACERC 1984). Long-period waves require a long fetch. The width of the fetch is also important, because, for fetches of the same length, a narrower body of water results in wave heights lower than those of open waters (Ippen 1966). Wave heights and wave periods are also lower if waves are generated in shallow, rather than deep, water. Wave heights of short-period waves are limited because they become unstable and break at lower heights than long-period waves (Komar 1976).

Johnson (1948) has described some characteristics of waves in lakes, and wind waves in deep-water reservoirs have been discussed and quantitatively analyzed by the Beach Erosion Board (1962). Low to moderate wave energy will generally characterize lakes of shallow water and limited fetch during non-storm conditions (e.g. Tanner 1971) but they also promote intense mixing and the development of currents by wind waves (e.g. Savkin 1975). However, even in shallow water and limited fetch lakes, high energy conditions are generated by low pressure storm systems and the greater wave intensity causes considerable, rapid changes to beach and bluff zones (e.g. Davis and Fox 1972b, 1975). Tables and graphs are available for estimating the significant wave height, H_s, and peak energy, based upon fetch and wind speed, as well as for estimating other parameters for engineering purposes (e.g. USACERC 1984). Vincent (1981) and Grosskopf and Vincent (1982) discuss approximate methods of estimating wind wave energy and nearshore significant wave height in shallow water.

As waves approach the shore, they are increasingly affected by bottom topography and configuration until the water depth shallows to approximately one-half their deep-water wavelength. At this point, their wavelength and velocity decrease while their height increases (e.g. Eagleson 1956, Koh and Le Mehauté 1966). Changes in wave profile also result from disruption of the orbital motion within the wave (Fig. 8). The wave form is distorted by interference with shallower foreshore regions and becomes more pointed and exaggerated in steepness closer to the land. Eventually the waves are oversteepened and break (Wiegel 1964, Collins 1976, Le Mehauté 1976). Waves break at the approximate point where their height equals the water depth. Breaking occurs because the velocity of water in the crest exceeds the phase velocity of the wave form (the rate at which the crest is moving forward), thereby becoming oversteepened (e.g. Wiegel 1964, Galvin 1968, 1972, Collins 1976, Cokelet 1977).

Four types of breaking waves are generally recognized, although in fact they are gradational in form from one to another and vary in form with beach slope and wave steepness (Fig. 9). Spilling breakers gradually peak until the crest becomes unstable and cascades down as turbulent "white

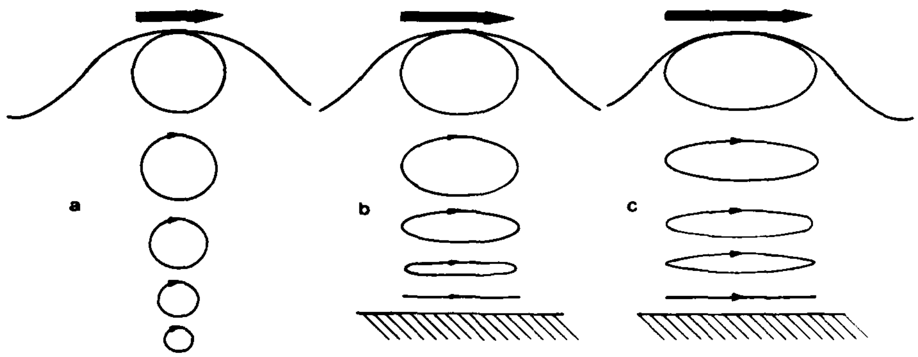


Figure 8. Orbital motion of water particles beneath surface waves in different depths of water. (a) deep water, (b) water of intermediate depth, (c) shallow water (after Allen 1982).

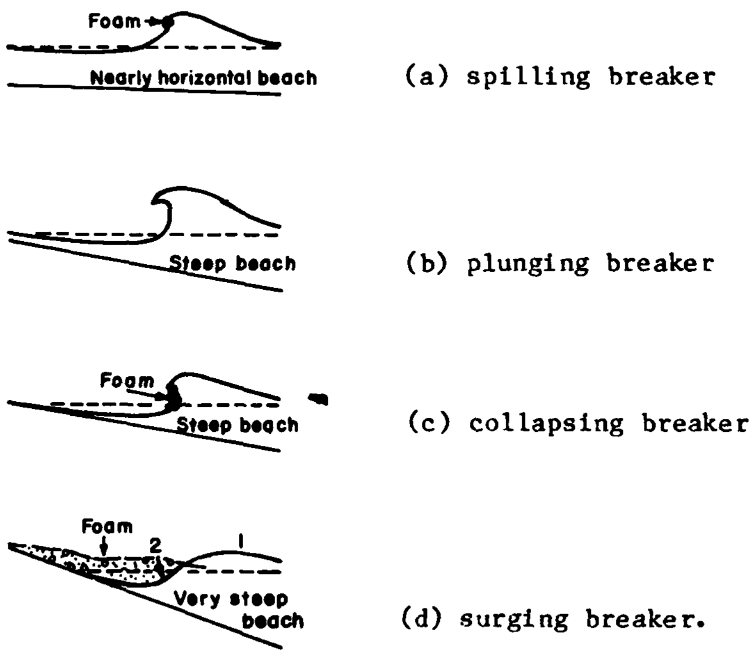


Figure 9. Four types of breaking waves.

water" (bubbles and foam). In plunging breakers, the shoreward face of the wave becomes vertical, curls over and plunges downward, striking the surface as an intact mass of water. Surging breakers collapse downward, the wave surging up the beach face. Collapsing breakers, intermediate between the plunging and surging types, result when the wave overturns below the crest within the forward face, the location being marked by white water (Galvin 1968, Cokelet 1977).

Breaker type depends upon the initial wave energy flux and thus offshore wave steepness (H/L), and on the rate of energy flux input from shoaling and, hence, the beach slope (Cokelet 1977). Plunging breakers

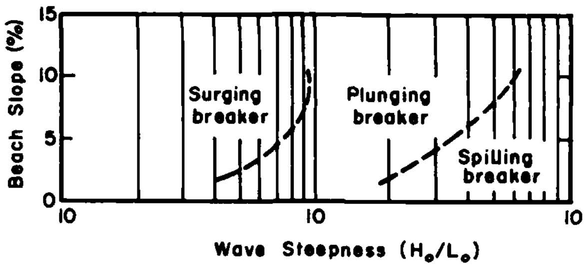


Figure 10. Type of breaking wave as a function of wave steepness and beach slope (after Wiegel 1964).

tend to occur on steeper beaches from waves of intermediate steepness, surging breakers occur on high-gradient beaches with waves of low steepness, and spilling waves occur on beaches of very low slope from waves of high steepness (Wiegel 1964) (Fig. 10). On a steep beach, waves steepen rapidly as they approach, and when plunging breakers develop, they collapse and dissipate energy in the narrow, turbulent surf zone within which swash action is important (Huntley and Bowen 1975). On shallow beaches, breaker form develops more slowly and waves undergo a steady transformation to a steep, "bore-like" frontal wave that dissipates the energy over a much wider zone.

In a reservoir with rapidly deepening waters close to the shoreline (and other factors being equal), plunging and spilling breakers may be expected to be more common than on natural lakes. Under conditions most conducive to erosion, waves would impinge directly on the base of bluffs and all wave energy would be dissipated on these sediments. Turbulence, eddies and other complex water movement could be expected in the waters adjacent to the bluff.

Sediment entrainment and transport by waves

Sediment transport in the offshore zone is induced by the orbital motions of waves as a combination of bed load and suspended load (e.g. Inman and Bowen 1963, Dingler and Inman 1976) (Fig. 11). The initiation of sediment motion and the depth to which waves may cause this to take place is dependent upon wave size and period, and hence maximum effective fetch and sediment properties (e.g. Hakanson 1977). Johnson (1981) estimated the depth to which waves will initiate motion on lake beds as a function of effective fetch, grain size and wind velocity (Fig. 12). Johnson's (1981) calculations compare reasonably well with field data gathered by Hakanson (1977) (Fig. 13). Depths of motion induced by waves of a particular period can be significant; as an extreme for ocean waves with a period of T = 15 s, depths of 100 m or more are possible (Komar and Miller 1975a). Nearshore wind wave-induced, unidirectional currents may further add to, or subtract from, the net offshore transport rate.

Inside the surf zone, turbulence due to breaking waves suspends bed materials and initiates motion of sediments on the bed (e.g. Eagleson 1959, Eagleson and Dean 1961) (Fig. 14), but the mechanics of the entrainment process are only partly understood (Miller 1976). Transport takes place both as bed load concentrated in a thin layer close to the bed and as sus-

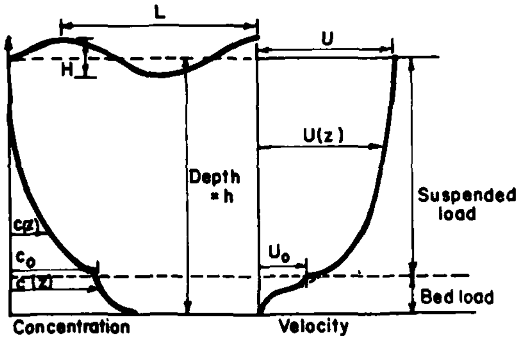


Figure 11. Relative concentration (c) of sediment in suspended load and bed load and orbital velocity (U) as a function of depth beneath a wave (after Muir Wood and Fleming 1981, p. 129).

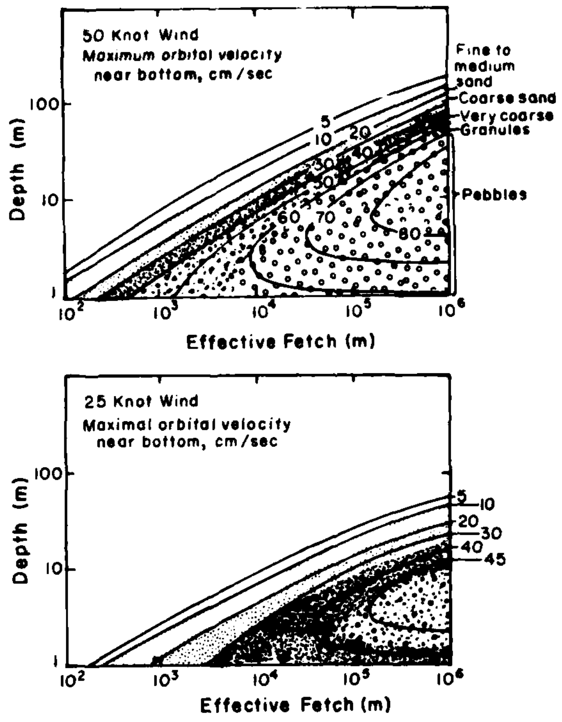


Figure 12. Calculated estimates of maximum orbital velocity at the lake floor as a function of effective fetch, depth and wind velocity. Maximum grain size of bed material which is moved at the estimated velocity is also estimated (from Johnson 1981, p. 1314).

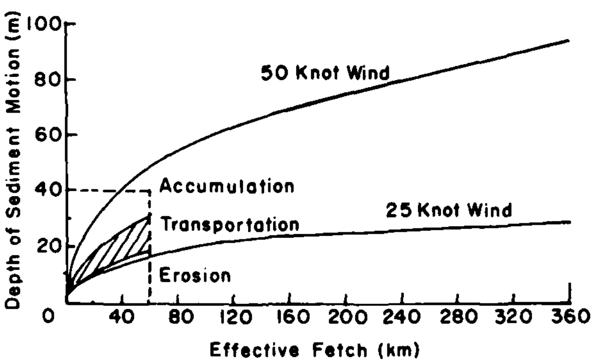


Figure 13. Comparison of field data of Hakanson (1977) for Lake Vänern (dashed box) for erosion, transport or deposition of sediment of 0.1-mm diameter at a given depth, with calculated values of Johnson (1981) for winds of 25 and 50 knots (after Johnson 1981).

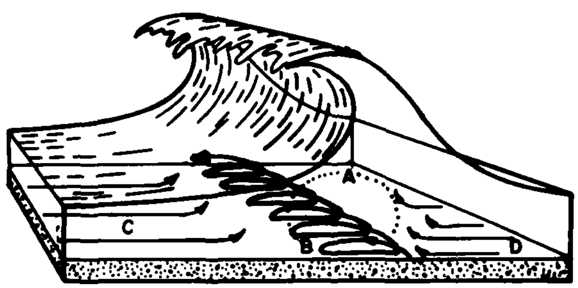


Figure 14. Schematic representation of particle motions in a breaking wave, with A representing suspended grains, B bed material, C grains shoreward of the wave, and D grains seaward of the wave (after Ingle 1966, p. 53).

pended load within the water column. Komar (1976) states that bed load transport dominates over suspended load transport on beaches, and also that only material which is coarse enough to remain in bed load will remain on the beach. Suspended fine-grained particles get transported from the beach in nearshore currents and thus silt and finer particles are absent from equilibrium beaches. Studies by Thomas et al. (1972) support this latter statement.

Although a rigorous mathematical treatment of the initiation of sediment motion by waves (or waves and currents combined) is lacking, empirical relationships have been proposed to estimate it. Komar and Miller (1973) concluded that, for grains of cohesionless material less than 0.5 mm in diameter, sediment movement is estimated by

$$\frac{\rho u_t^2}{(\rho_s - \rho)\, g\, D} = 0.21 \left(\frac{d_o}{D}\right)^{1/2} \qquad (1)$$

where u_t = the <u>critical velocity</u> to initiate motion
d_o = diameter of the orbital wave motion
ρ = water density
ρ_s = grain density
D = grain diameter
g = acceleration of gravity

The critical velocity and wave orbital diameter are related by the following relationship:

$$u_t = \frac{\pi\, d_o}{T} = \frac{\pi H}{T \sinh (2\pi h/L)} \qquad (2)$$

where

H = wave height
h = water depth
L = wave length
T = wave period.

For coarse sand and larger grain sizes (> 0.5 mm diameter), the empirical relationship is

$$\frac{\rho u_t^2}{(\rho_s - \rho)\, g\, D} = 0.46\, \pi \left(\frac{d_o}{D}\right)^{1/4} . \qquad (3)$$

Estimates can thus be made if the average grain diameter and density, wave period T, and either orbital velocity or orbital diameter are known. Komar and Miller (1975a,b) evaluated the sediment mobilization threshold under various wave conditions, including those shown on graphs based upon eq 1 and 3 (Fig. 15),and developed a computer program to define sediment threshold for field applications. For a particular grain diameter, several combinations of T and u_t are possible: for longer values of T, the greater the u_t required. As indicated by the relationship for u_t in eq 2, many combinations of water depth and wave height can produce the required u_t.

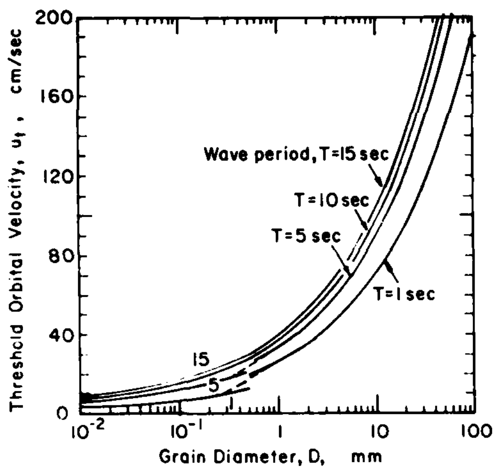


Figure 15. Threshold of sediment motion by waves estimated for cohesionless material of a given diameter D and density of 2.65 g/cm^3 (quartz) (from Komar and Miller 1975a). Equations in text define relationship to wave height and water depth.

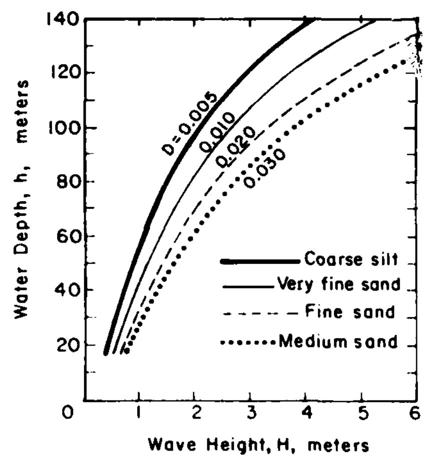


Figure 16. Water depth at which sediments are mobilized by surface waves of period T = 15 seconds (from Komar and Miller 1975a).

Komar and Miller (1975a) and Madsen and Grant (1976) also concluded that the initiation of movement of cohesionless particles under oscillatory unsteady flow could be estimated with the empirical relationship developed by Shields (1936) for movement under unidirectional steady flow in shallow water as the result of a critical tractive force exerted by currents at the bed. The maximum bottom shear stress for oscillatory motion must be determined in order to apply the Shields criterion, a value which has not yet been measured. The concepts of Shields (1936) and others are discussed in the section entitled Overland Flow.

Grant and Madsen (1979) developed a theory for calculating maximum bottom shear stress when both waves and currents interact, again estimating the onset of sediment motion with Shields' (1936) empirical relationship. Their theory suggested that when waves and currents combine, the resulting bottom shear stresses are altered by turbulence at the bed interface and differ from the stress expected for either waves or currents alone. A resistance to flow essentially results that is based upon an apparent bottom roughness factor. Turbulence at the bed is greater when currents move in the same direction as the wind waves (Kemp and Simons 1982), which increases the bed shear stress while decreasing current velocities. Additional flume studies by Kemp and Simons (1983) suggested that waves propogating against a current had similar effects on turbulence and velocity near the bed.

The depth to which waves initiate sediment movement can also be estimated with eq 1 and 3 (Komar and Miller 1975a). Figure 16 illustrates this effect for different grain sizes of bed material assuming a wave generated in an ocean with a period of 15 s.

As with all proposed theories on the threshold for sediment movement, limitations exist and result in inaccuracies. The relationships of Komar and Miller (1975a) give conservative estimates. Wave motion was assumed to be sinusoidal, but natural waves vary within any particular wave spectrum and have orbital velocities and motions that likewise may vary. Bed roughness also affects entrainment and is not always considered by these relationships. A more rigorous discussion of critical shear stress and sediment transport under wave action is given by Muir Wood and Fleming (1981).

The concentration of suspended sediment in waves at their breaking point is high, with the amount varying mainly with breaker type (Fairchild 1972, Kana 1978, 1979). Internally, sediment concentration decreases exponentially above the bed, with coarse bed material intermittently entrained in water near the bed under certain wave conditions (Fig. 11) (Muir Wood and Fleming 1981). Kana (1979) analyzed natural waves and found that plunging breakers entrain one order of magnitude more sediment than spilling breakers which suspend very little material. Plunging breakers are particularly important in scouring beach sediments because they develop large, near-vertical velocity vectors at impact (e.g Adeyemo 1971, Cokelet 1977, Allen 1982a). Kana (1979) also concluded that factors determining sediment concentration included beach slope, wave height, and distance from break point, but the concentration is independent of wave period, longshore current velocity, wind velocity or wave steepness. The efficacy of waves to scour and entrain sediment may also be affected by water temperature because of the viscosity change, but for turbulent plunging breakers, this effect would be negligible in comparison to the effect of turbulence.

Sediments scoured by plunging breakers are normally moved as part of the swash and backswash of water across the foreshore due to the inertia from the breaking waves. Suspended fines and coarser particles rolled up the beach face may be entrained within nearshore currents which are also generated by the waves. Swash and backswash can reach heights on the foreshore significantly above the water level.

During breaking and run-up of waves, distinct and significant variations take place in the water table of the beach sediments (e.g. Grant 1948, Waddell 1973, 1976, Chappell et al. 1979, Kondratjev 1966). Under swash and backswash due to wave action, oscillations in groundwater create a zone that is periodically saturated or unsaturated. Under unsaturated conditions, water flows into beach sediments and localized deposition is favored. When saturated, infiltration cannot occur and previously deposited sediment can be eroded. This is particularly important on sandy or coarser beaches because infiltration adds to the effects of gravity and friction in causing cessation of the swash motion. Kondratjev (1966) suggested that pressure variations on the bottom due to oscillatory wave motion can induce water flow into pore spaces and energy dissipation. In coarse-grained gravel beaches, he estimated inflow to pore spaces to absorb half the available energy and used this concept to explain the lower angles of slope in run-up areas of beaches composed of finer-grained material and the higher slope angles on coarse-grained beaches.

Chappell et al. (1979) concluded that rises in the water table in sediments beneath and landward of the beach face, which are covered by steadily rising water levels from tides, takes place as a slow wave of diminish-

ing amplitude and increasing time lag. Under a rising water table, pressure waves propagating into the water table from breaking waves can induce slumping by initiating liquefaction of the sand. Lowering the water table induces sand deposition and also reduces the tendency for such liquefaction. Chappell et al. concluded that this factor is important in beach failure and slumping when water levels rise during storms. Similar effects are possible in reservoirs when pool levels fluctuate; this may account for bank failures following water level increases that are coincident with intense wave activity.

NEARSHORE CURRENTS

Waves entering shallow water undergo refraction such that the direction of travel approximately parallels the depth contours of the nearshore zone (Fig. 17). Crests become nearly parallel to the shoreline by the time the waves reach the beach (e.g. Bretschneider 1966). In many "young" reservoirs, however, deep waters lie close to the shoreline and waves may impinge on the beach at angles widely divergent from the trend in the shoreline. This phenomenon is important because wave-induced longshore currents, which transport sediment mobilized by waves to replenish and maintain down-current beaches, are produced by oblique wave approach (Komar and Inman 1970, Longuet-Higgins 1970a,b).

Irregular bottom topography, often characteristic of reservoirs, is also important because variation in wave height and the distribution of wave energy along the shoreline can result from complex refraction and reflection patterns (e.g. Munk and Traylor 1947). Waves refract and flow lines diverge over deeper water areas whereas they converge with movement over shallows (Fig. 18). Wave energy is greater in the areas of convergence because of the resultant increase in wave height and it diminishes in areas of divergence shoreward of, or adjacent to, shallows where wave height is reduced. Intense areas of erosion and shoreline recession may thus be correlated to wave convergence over nearshore shallows (e.g. Maresca 1975). Hence, shoreline morphology is partially controlled by offshore bathymetry (Fico 1978) (Fig. 19).

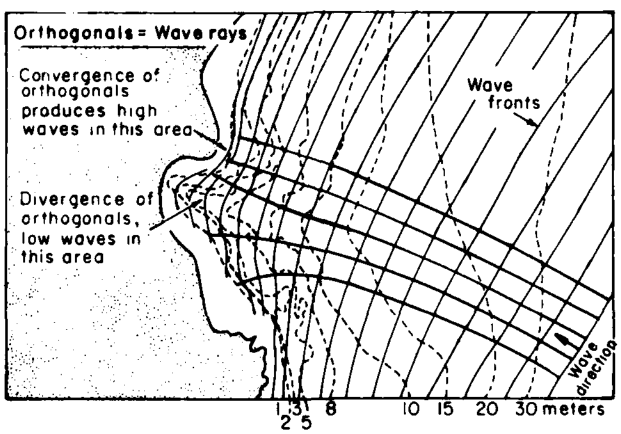


Figure 17. Idealized relationship between monochromatic waves, depth contours and shoreline configuration (after Goldsmith 1976).

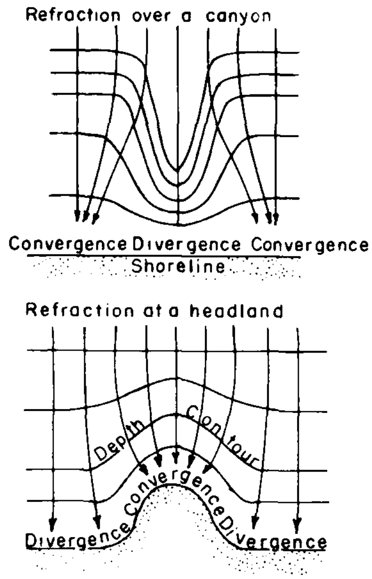
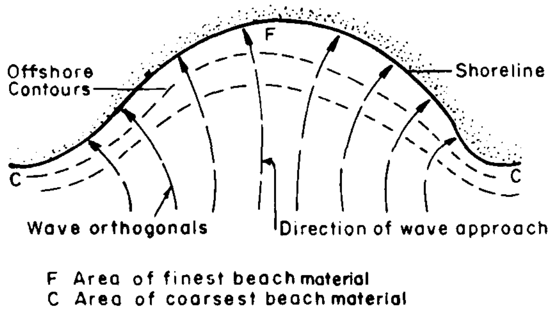
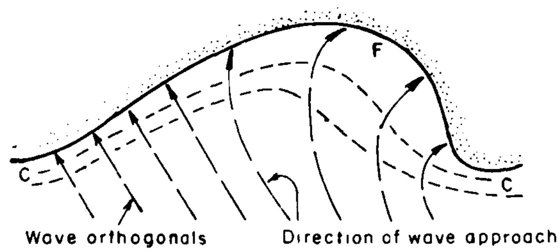


Figure 18. Refraction of wave crests in response to changes in water depth near the shoreline. Wave energy is greatest in areas of wave ray convergence; least in areas of divergence (after Komar 1976).



a. Bay facing directly into prevailing winds



b. Bay facing obliquely into prevailing winds.

Figure 19. Bay form and shoreline configuration in relation to prevailing waves (after Muir Wood and Fleming 1981). Mean grain size of beach material reflects differences in wave intensity with convergence or divergence.

Wave-induced currents, called the nearshore cell circulation system, predominantly control water movement in the littoral zone (Fig. 20). This system consists of rip currents that are fed by longshore currents and develops in response to mainly longshore variations in wave height or wave set-up (Shepard and Inman 1950, Bowen 1969a, Bowen and Inman 1969).

Because of irregularities in bottom topography, water shoreward of higher breakers is raised above that shoreward of smaller breakers, causing water to flow toward the area of smaller waves. These flows converge and

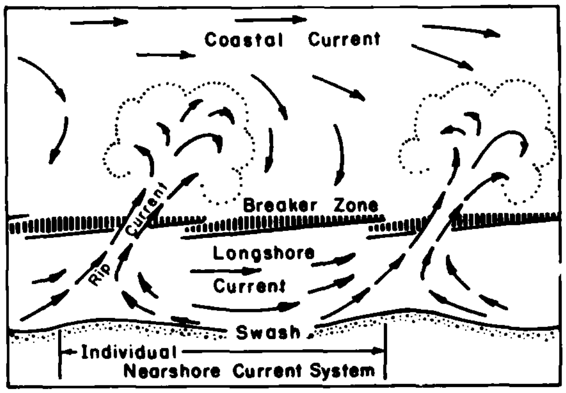


Figure 20. Diagram of nearshore current system illustrating the wave-induced longshore and rip currents along a shore with a protuberance to the shoreline (after Komar 1971).

move seaward as a single rip current. Variations in wave height, due to wave refraction (Shephard and Inman 1950) or due to the interaction of edge waves and incoming waves in the nearshore region (Bowen and Inman 1969), may also result in a cell system. After development, nearshore bars or troughs previously scoured by rip currents may also channelize nearshore waters into cell circulation systems, even when wave height or edge wave variations are minimal (Sonu 1972).

Longshore currents are produced by the oblique approach of waves and their striking of the beach at an angle to the shoreline trend (Bowen 1969b, Longuet-Higgins 1970a,b, Komar and Inman 1970, Thornton 1971). The interaction of these obliquely traveling waves with a variable longshore wave height results in an asymmetrical current pattern, with longshore currents feeding rip currents (Komar and Inman 1970, Komar 1975, Fig. 20). Observations suggest the entire nearshore cell system may migrate along the shore (Bowen and Inman 1969), and thus the nearshore features and shoreline configuration will also migrate (Sonu 1973). Both longshore and rip currents are affected by shoreline configuration and shallow water topography.

Longshore current generation strongly depends upon wave height and wave phase speed (celerity). Waves in lakes often have short periods and wavelengths and are rather steep, and they are therefore not strongly refracted near the shore (Davis 1976b); consequently, they can generate rapid along-shore currents and with them, significant sediment transport.

Sediment transport by currents

Of the nearshore currents, longshore currents are the primary process transporting sediment in the littoral zone. Because longshore currents are a component of the wave system, sediment transport varies with wave energy, wave steepness and angle of wave approach (Inman and Bagnold 1963). In a general sense, total net sand transport in the littoral zone results from the orbital velocity of incoming waves (which places beach sediments in motion) and the longshore, rip and currents normal to the shoreline (such as those generated locally by the wind), which transport it along the shoreline or directly into offshore regions (Bagnold 1963, Komar and Inman 1970, Komar 1971). Total net sand transport is approximately proportional to wind stress and current velocity, and thus the rate of movement is at a maximum nearest the breaker line, decreasing shoreward. Sediment movement is enhanced when currents and wave motions are combined (e.g. Bijker et al. 1976).

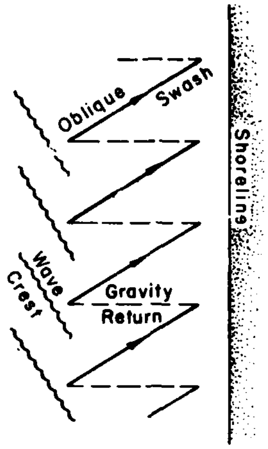


Figure 21. Conceptualized zigzag motion of sediment along a beach face under wave swash. The incoming wave moves sand at an oblique angle to the shoreline and the backswash flow due to gravity moves it back to its original level on the beach (after Komar 1976).

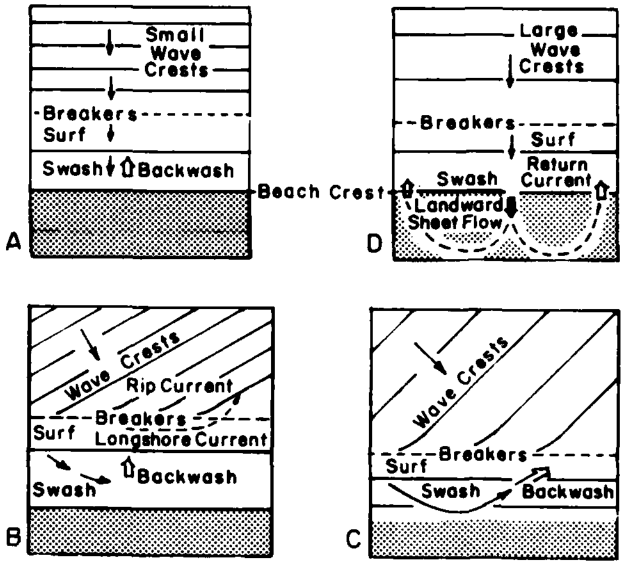


Figure 22. Schematic of variations in swash and backswash motions of particles due to head-on approach of small (A) and large (D) waves, and oblique approach of small (B) and large (C) waves. Large waves associated with storms create a vigorous swash and backswash involving large volumes of water with significant erosion (after Friedman and Sanders 1978).

This movement by combined waves and currents has been described by Komar (1971) as a zig-zag motion (Fig. 21). Each incoming wave drives sand particles up the beach at an oblique angle determined by the angle of the approaching wave. A return flow or backswash results from gravity effects and moves the particle back to its original level on the shore profile. Depending upon angle of approach and wave characteristics, the swash-backswash motion of particles may differ (Fig. 22).

According to Bagnold's (1963) model, the orbital motion of waves moves sediment back and forth after initially suspending it within the water column. Net transport, however, only results from the unidirectional currents that are present. Thus,

$$i_\theta = K' \omega \frac{u_\theta}{u_o} \qquad (4)$$

indicates the sediment transport per unit width i_θ occurring in the direction θ under a current u_θ. u_o is orbital velocity, K' a dimensionless parameter, and ω represents the force exerted by the waves (Fig. 23). Thus ω/u_o is the stress exerted by the waves.

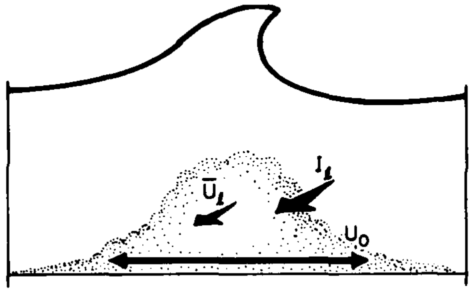


Figure 23. Schematic representation of net sand transport by waves I_ℓ as the result of the orbital velocity u_0 of the waves placing particles in motion and the current u_ℓ transporting them (after Komar 1976).

In order to estimate total sediment transport for breaking waves, Inman and Bagnold (1963) applied this relationship by assuming that the wave energy flux per wave crest (defined as P cos α) is lost in initiating sediment motion. Thus the stress applied to the beach sediments is now proportional to P cos α/u_{os}, where u_{os} is wave motion speed relative to the bed and is proportional to u_0 before the wave breaks in the surf zone. For a location with a longshore current of velocity v_ℓ, total transport I_ℓ (e.g. Komar 1976) is as follows:

$$I_\ell = K' \text{ (ECn) } \cos \alpha \frac{v_\ell}{u_{os}} \tag{5}$$

where P = the standard wave power function = ECn
E = total wave energy density
C = wave phase velocity
n = a ratio of wave group and wave phase velocities
α = the angle between the breaking wave crest and the shoreline.

Studies by Komar and Inman (1970) indicated that for K' values of 0.28 and $u_{os} = u_m$ (the maximum orbital velocity under the breaking wave), eq 6 gave a reasonably good prediction of littoral sediment transport for wave and current interaction. By using wave parameters based on hindcasting techniques of waves generated by storms, littoral transport rates at a location can be grossly estimated (e.g. USACERC 1984).

Onshore/offshore sediment movement

Perhaps of more importance to erosional changes in the shoreline is the interrelationship of wave-induced oscillatory motions, unidirectional currents, onshore-offshore sediment movement, and the resultant changes in equilibrium beach profiles. McGreal (1979), for example, analyzed various environmental factors, concluding that short-lived, temporal variations in erosion rates were largely explained by changes in beach-profile configuration and elevation, while along-shore changes were related to the aspect of the shoreline sites. Weischar and Wood (1983) found that longer term shifts in beach profile could be correlated directly with wind and water level changes on the Great Lakes.

Net movement or shifts in sediment from onshore to offshore regions have been studied from the standpoint of seasonal variations in shore profiles associated with seasonal changes in storm activity and thus wave intensity (e.g. Shepard 1950b, Bascom 1954, Aubrey 1979), and as the result

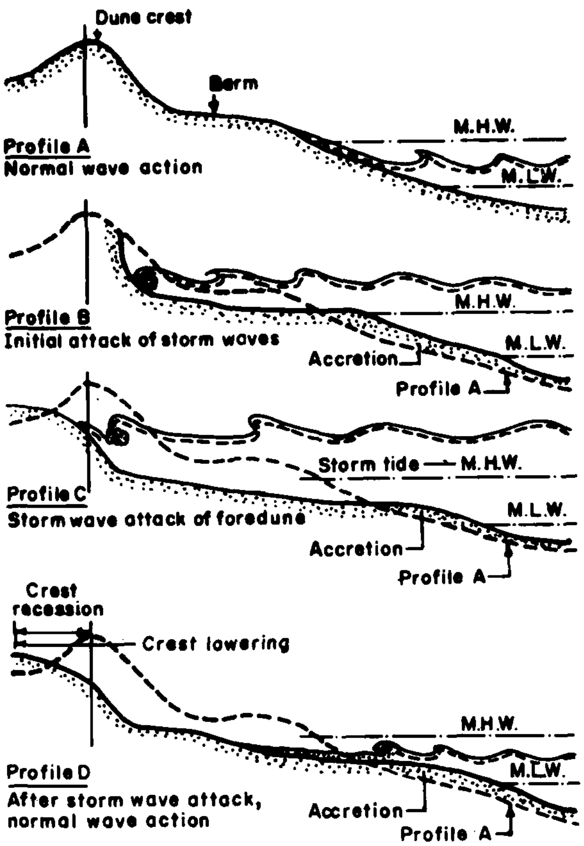


Figure 24. Idealized changes in shore profile resulting from a single storm-generated wave attack (after USACERC 1984). (MHW-mean high water level; MLW-mean low water level).

of intense wave activity during a single storm (e.g. Davis et al. 1972, Hayes and Boothroyd 1969, Fig. 24). While the summer and winter profiles resulting from annual variations in shore processes last several months, shifts from a storm profile to those characteristic of calmer conditions (swell profile) may begin almost immediately following the passage of the storm. A critical wave steepness apparently exists that can be related to shifts from swell to storm profiles (Iwagaki and Noda 1963), but a precise means of predicting this value is not available (e.g. Rector 1954, Watts 1954, Kemp 1961, Dean 1973 in Komar 1976, Komar 1976). Other parameters clearly related to onshore or offshore shifts in sediment are wave height and possibly wave period, but results thus far are inconclusive (e.g. Shepard and LaFond 1940, Shepard 1950a,b, Bascom 1954, Gorsline 1966). Onshore-offshore shifts in sediment may also result from coastal winds (e.g. King and Williams 1949, Shepard and LaFond 1940, King 1953, Seibold 1963, Weishar and Wood 1983) and tides (e.g. Strahler 1966, Duncan 1964). Tidal effects may provide conditions analogous to water level fluctuations in reservoirs, since they occur on an hourly scale.

Offshore moving sediment in lakes and coastal areas is usually deposited in one or more longshore bars that have troughs on their shoreward side (e.g. Evans 1940, Shepard 1950a, King and Williams 1949, Davis et al. 1972, Greenwood and Davidson-Arnott 1979). Bars migrate onshore during non-storm periods (e.g. Davis and Fox 1972a). Their location and size are controlled mainly by breaker position, wave height and wave steepness (e.g Keulegan 1948, Felder and Fisher 1980). Plunging breakers appear most conducive to bar development (Shepard 1950a), with the deeper waves generating larger bars (Keulegan 1948).

The mechanics and factors controlling onshore or offshore sediment movement are complex and, to date, defy full theoretical treatment. Profile changes are the result of the interaction of shallow water waves and wind- and wave-induced currents. Bagnold (1963) modeled such an interaction for the region inside the breaker zone as orbital wave motions that suspended sediments, upon which are superimposed the unidirectional currents of the littoral zone that actually transport sediment. Inman and Bowen (1963) examined such interactions in deeper water beyond breakers. Their study indicates that several parameters can affect whether a net onshore or offshore movement takes place.

Experimental and theoretical treatments by Ippen and Eagleson (1955) and Eagleson and Dean (1961) attempted analysis of when particles in the onshore or offshore regions may be expected to move and the direction of that movement. Komar (1976) points out that such treatments lacked realistic depiction of natural beaches because they ignored the effects of bedforms on flow. Recently, Quick (1983) discussed the interaction of sediment transport by breaking waves and the effects of currents moving in the same or opposite directions as the waves. His measurements indicated that transport is determined by the direction of wave motion until waves start to break. Thereafter adverse currents cause reversal of the transport direction.

The net movement of sediments from eroding bluffs and beaches of reservoirs is often assumed to be in the offshore direction because the nearshore zone and profile are not considered to be in equilibrium with the lacustrine conditions (e.g. Kondratjev 1966). Nearshore beach slope angles are generally steep and retard or inhibit the onshore and longshore movement of sand. Consequently, eroded material is not replenished and nearshore areas will be characterized by only offshore sediment movement until erosional and/or accretionary changes develop a shore zone profile that is in equilibrium with the reservoir pool processes. Then, overall adjustments in sediment distribution, shoreline configuration and the shore profile will be similar to those of coasts and lakes.

Shore zone profile changes

The concepts of net onshore-offshore sediment movement have been used by authors of conceptual models to account for erosion and shoreline modifications. Bruun (1962) formulated a conceptual model for profile adjustments to onshore/offshore sediment-migration due to a rising sea level (Fig. 25). Bruun suggested that as sea level rises, material will be eroded from the upper foreshore and backshore zones and redeposited offshore. The net result is a shoreward displacement of the beach profile. A necessary assumption of this model is that the shore zone profile is in a dynam-

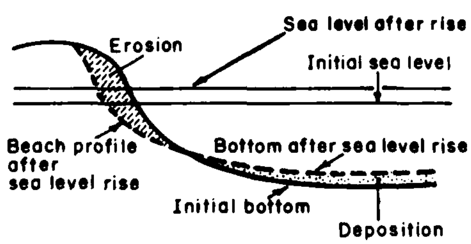


Figure 25. Conceptual model of Bruun (1962) for profile adjustments resulting from a rising sea level.

ic equilibrium with water levels before the rise begins. Schwartz (1965, 1967) and DuBois (1975, 1976, 1977) have presented laboratory and field evidence generally supporting and refining Bruun's concept of the effects of rising water level on beach erosion.

Field studies by Moody (1964) provide some indication of the processes by which such shifts in sediment and changes in the shore profile take place. Moody's study points out the importance of large storms in generating high intensity wave energy and rip currents that transport shore materials into the offshore zone. Coupled with the offshore bottom flow associated with the wind tide, such storms result in rapid and large changes in the beach and bluff zones over relatively short time periods (hours). It is of course inherent in the concept of profile changes that a lowering of water level will result in deposition and a seaward translation of the shore profile (e.g. Curray 1964, Swift 1976). Locations where net onshore movement of sediment takes place will show a similar accretion of materials.

Hands (1980) has applied Bruun's concept to erosional changes of the shore zone along the Great Lakes and suggests that under the conditions of rising lake levels, this concept can be used to predict net long-term changes in shore profiles. Hands (1980) stresses, however, that the extreme simplicity of the model and the intractable complexities of actual beach and nearshore processes require its careful application only after a detailed assessment of field conditions. Hand's predictive model will be discussed in more detail in <u>Models of Shore Zone Development and Erosion</u>.

SUMMARY OF WAVE AND CURRENT EROSION

Net erosion by wind waves depends primarily upon the following factors: 1) wind velocity, duration and effective fetch, 2) nearshore and offshore bathymetry, 3) shoreline configuration, 4) water level, and 5) beach and bluff composition (Kachugin 1966, Wiebe and Drennan, 1973, Edil and Vallejo 1980). Because these factors can vary widely between reservoirs as well as within a single impoundment, the occurrence or rates of erosion by wind waves can vary greatly along adjacent lengths of shore.

In general, the effect of wind waves on beach and bluff zone erosion in reservoirs is two-fold. First, direct wave attack results from waves breaking directly upon the backshore or bluff sediments. The actual erosion apparently results from a variety of forces exerted by the wave within the sediments and an abrasive action by the particles entrained within the wave (e.g. Sunamura 1977, Robinson 1977, McGreal 1979). Cyclic loading by waves on undrained, saturated beach sediments can progressively increase pore pressures, thus reducing their resistance to shear (e.g. Seed and Rahman 1977) and increasing their erodibility by wave turbulence or currents.

Secondly, swash run-up following wave breaking may cause erosion, although much of the wave energy is lost after breaking due to percolation and frictional effects of swash movement. Abrasion by entrained sediment and hydraulic erosion and transport may still result; its effectiveness will depend upon beach and bluff composition (McGreal 1979). Swash run-up is perhaps more important in entraining sediment at the bluff's toe that

had been previously eroded and deposited from the bluff face by other processes (e.g. Edil and Vallejo 1980).

Removal of sediments at the bluff's base can prevent a lowering of the bluff/beach slope angle and thus the establishment of an equilibrium shore profile. Sediment eroded and entrained by wave-generated currents is added to the littoral transport system and moved either off or along the shore. Deposition of this eroded sediment in nearshore waters is important in developing the equilibrium profile. Offshore bathymetry determines wave refraction patterns and hence the alongshore variability in wave energy impinging on beaches and bluffs and the direction of movement of eroded material by longshore and rip currents. Temporal changes in the location of wave attack due to atmospheric or artificially induced changes in water level will modify the locations and intensity of attack by breaking waves and thus the locations of sedimentation (e.g. Maresca 1975, McGreal 1979, Carter et al. 1981).

STORMS AND SEASONALITY

Although the effects of storms are strongly correlated to wind waves and nearshore currents, I have singled them out in this report because of their potential importance as events that can cause rapid and extensive modification of the shore zone over short time intervals. Significant changes in shoreline position commonly result from major storms (e.g. Fox and Davis 1970, 1976, Trepetsov 1972, Maresca 1975, Savkin 1975, Davis 1976, Hamblin 1976, Cogley and McCann 1976, Abele 1977, Goldsmith et al. 1977, Birkemeier 1981, Vellinga 1982); some of these authors concluded that without severe storms, erosion and shoreline recession would be almost nonexistent under present conditions.

Storms produce beach and bluff erosion mainly because of the strong winds and the resultant increase in wind stress that generate larger waves. Wave propagation and wave energy expended on the shore zone increase. Both breaker height and longshore current velocity apparently increase as the barometric pressure drops with passage of low pressure systems in Lake Michigan (Davis and Fox 1972b). Heavy downdraft winds along a squall line thunderstorm can cause large seiches and move the wave attack higher on beach or bluff sediments (Hunt 1959, Hamblin 1976). On Lake Michigan, recession of bluffs is rapid during storms but only when waves break at or above the normal shoreline (Maresca 1975). Transport of erosional products is mainly offshore by rip currents fed by longshore currents on the downwind side of storms in this lake, but passage of the storm often reverses both the wind and current direction. Offshore bar dimensions and locations also change rapidly during storm passage.

An additional effect observed on coasts is that waves generated by intense storms can cause rapid changes in the effective stress of bottom sediments and increased pore pressures, and under undrained conditions, lead to loss of structure and reduced strength or remolding of these materials (Henkel 1970, Sukayada et al. 1976). The stability of subaqueous slope sediments may be reduced sufficiently to induce failure and mass movement. Theoretical analyses by Moshagen and Torum (1975), however, suggested that pressure gradients will only be important in fine-grained bed material to some limited depth. Posey (1971), in contrast, concluded that the excess

pore pressures can be sufficient to increase bed erodibility by nearshore currents.

The effects of storms are dependent upon several factors that include their intensity, duration and frequency of occurrence, offshore and foreshore topography, water depth, height, orientation and configuration of the bluff, and composition of bluff and beach materials (Seibel 1972, Maresca 1975, Savkin 1975, Davis 1976a,b, Fox and Davis 1976, Birkemeier 1981). Alongshore variability in the rate of erosion and shoreline recession results primarily from variations in the shore zone's orientation, composition, topography and proximity to the storm's path (Birkemeier 1981).

Maresca (1975) concluded that storm frequency is important because removal of beach material by a storm, without its replenishment by normal waves and longshore currents, reduces the protection of the beach. The smaller runup area or absence of a beach after several storms allows waves to attack bluffs without loss of energy due to runup. This situation is one analogous to that of newly formed shore zones of reservoirs that are without a beach and have deep water directly adjacent to the shoreline. In these reservoirs, direct attack of the bluffs occurs without the replenishment of the eroded sediments found on beaches that exhibit an apparent dynamic equilibrium (e.g. Davis and Fox 1972a, Fox and Davis 1978).

Maresca (1975) also observed that blufflines on Lake Michigan receded at maximum rates at distinct distance intervals along the shoreline. These intervals were 1) a spacing of about 1 km, over which long sections of shoreline had severe recession that correlated with wave energy distribution as affected by wave refraction, with recession rates highest at zones of wave convergence during storms, 2) a spacing of 100 to 200 m, over which sub-areas of bluff recession correlated with rhythmic changes in the bluffline, shoreline and offshore topography, and 3) a spacing of 10 to 30 m, over which recession varied because of large-scale bluff failures.

Seasonal variability in erosion and recession rates can often be related to the seasonal changes in storm frequency and intensity (e.g. Davis 1976a, Abele 1977, Fox and Davis 1978). Seasonality influences erosion through 1) changes in storm intensity and duration, 2) changes in water and air temperatures, 3) presence or absence of an ice cover, 4) presence or absence of a snow cover, and 5) timing of spring breakup and runoff.

Several authors working on the Great Lakes and coastal areas observed that breaking wave height and period increased from late spring to winter, with the highest waves from January to March (e.g. Fox and Davis 1978, Birkemeier 1981). An increase in such seasonal storms can lead to the repetitive removal of beach sediments as discussed by Maresca (1975). In general, seasonal changes in beach character result from removal of finer beach sediments into the offshore zone, these materials being returned to the beach surface in the spring as the wave climate changes (e.g. Shepard 1950b, Bascom 1954, Fox and Davis 1978).

In northern areas, this increase in intensity, frequency and duration of storms can cause rapid recession of bluffs, but the effects of such storms can be severely limited by the formation of an ice cover and freeze-up of bluff and beach sediments (Birkemeier 1981, Taylor 1981) (Fig. 26).

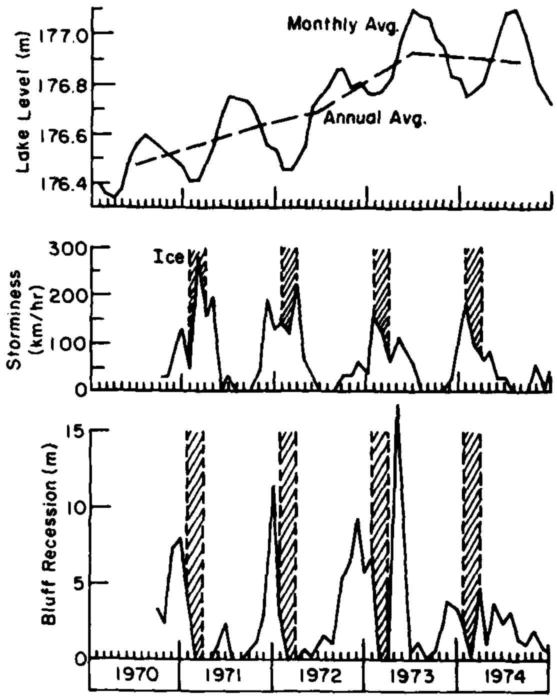


Figure 26. Relationships of changes in lake level, storminess and bluff recession over time on Lake Michigan (Birkemeier 1981). The effects of the ice cover are to protect the beaches and bluff toe from the direct attack of winter storm generated waves. Storminess is arbitrarily defined as sum of average daily wind speed between two shore profile surveys when onshore winds were greater than 29 km/hr.

This illustrates the importance of timing of both freeze-up and of spring breakup. A late freeze-up and ice cover formation or an early breakup could expose the beaches and bluffs to the force of the more severe winter storms.

WATER LEVEL FLUCTUATIONS

Artificial impoundments are characterized by fluctuations in water level that may occur hourly, daily, weekly, monthly, seasonally or even yearly. In addition, the initial filling of an impoundment results in a continuous gradual rise in water level over a longer period of time.

When the water level fluctuates, it may have several overall effects:

1. Up or down translation of subaqueous and subaerial erosional processes, with possible effects on their intensity and importance.

2. Modification of the ground water regime and slope stability.

3. Killing and removal of vegetation.

4. Modification of the ice cover, snow cover and ground frost during winter.

These effects can in turn influence beach and bluff erosion processes, resulting in significant modifications to the shoreline, and can cause various environmental changes that will affect the biology of the impoundment (e.g. Austin 1979, Ploskey 1982). The duration of time over which the

water remains at a high or low level will strongly influence the severity of each of the four effects listed above.

Lateral shifts

The raising of the water level relative to a fixed shoreline profile and configuration will at first result in a simple flooding of backshore sediments. With time, the rising waters will have exposed upper beach sediments and subsequently bluff sediments to the direct attack of waves and currents (e.g. Savkin 1975, Zaruba and Mencl 1976). As the water level rises, the distance over which wave runup occurs is shortened as well as the distance from shore at which waves break. Both cause an effective increase in wave energy (McGreal 1979, Carter et al. 1981). The height of the junction between the bluff base and the beach zone therefore determines whether the normal water level fluctuations due to reservoir operations cause bluff erosion at the high water level.

Wave attack can undercut reservoir bluffs and induce failure as the result of this support removal (e.g. Gatto 1982a, Reid 1983). Pezzetta and Moore (1978) found that while wave attacks accelerated the overall process of undermining and bluff failure, erosion still depended upon other processes including overland flow processes, frost action and water table fluctuations. Higher water levels that are concurrent with storm activity result in further acceleration of erosion rates (e.g. Coakley and Cho 1972).

Conversely, a lowering of the water level will increase the potential distance for wave runup and thus less potent breaking waves will impinge upon beach and foreshore sediments (Fig. 27). Small cliff or bluff faces cut into the beach materials may mark the shoreline at these quasi-stationary lower water levels. Subaerial processes can directly affect beach

Figure 27. Foreshore sediments exposed by a lowering of water level. Small erosional berms are formed by wave attack during transient standstill in level between the maximum and minimum levels.

sediments, particularly during periods of heavy runoff or freeze-thaw that occur when the water level is lowered. Bluff materials that are loosened and fall to the base of the bluff will remain there until the water level rises, and thus add some protection to the bluff (e.g. Reid 1984).

Many investigators have examined shoreline recession on the Great Lakes during and following the long-term rise in water level from 1967 to 1976 (e.g. Saylor and Hands 1970, Seibel 1972, Omohundro 1973, Hands 1976, 1979, Pezzetta and Moore 1978). A rise of 0.8 m in the annual mean lake level from 1967 to 1973 resulted in recession rates of 1 to 6 m/yr in study areas of Hands (1979), compared to a 120-yr average of about 0.43 m/yr. By fall of 1976, the accelerated erosion rate had fallen to lower levels and ceased in many areas as lake levels stabilized. Based upon his analyses, Hands (1979) presented a linear relationship between increases in water level and mean shore retreat of 4 m of retreat per 0.1 m of submergence during the 1967 to 1976 rise. He also noted, as have others, that the more recent period of falling water levels has caused accretion of shore materials and lakeward movement of the shoreline at some locations.

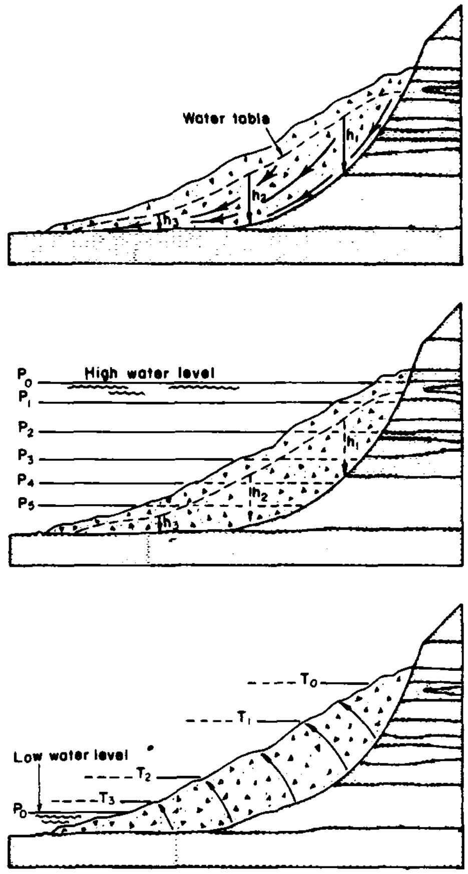
<u>Ground water changes</u>

As a reservoir is filled, the ground water table in the bank sediments gradually rises and adjusts to the water level in the impoundment. Water flows into bank sediments and, depending upon the hydraulic conductivity of the sediment, will eventually establish a new water table and develop pore pressures about equal to the hydrostatic pressure in the reservoir pool (e.g. Wahlstrom 1974, Fig. 28). This effect of reservoir water level on ground water conditions will diminish with distance from the shoreline (Virkulina 1977).

A ground water flow regime will thus establish itself, commensurate with (1) the stabilized water level following filling, (2) the composition of the bank sediments, and (3) the ground water conditions away from the immediate shore zone. Depending upon the porosity and permeability of the bank sediments, the water table will fluctuate in response to both external parameters (such as seasonal precipitation) and to water level variations in the reservoir (e.g. Jones et al. 1961, Simons and Rorabaugh 1971). The actual movement of ground water is extremely complex (Newlin and Rossier 1967).

The resultant pore pressures in bank sediments are mostly greater than before reservoir filling. Thus, the effective shear strength of the materials is reduced and, together with the increase in weight resulting from the water added to the bank sediments, can create an unstable situation that will lead to failure of those materials (e.g. Brunsden and Kesel 1973, Hopkins et al. 1975). The composition and stratigraphy of the bank sediments will also affect the failure condition, with the least stable strata or material controlling overall bank stability in most instances. Combining near-failure conditions, due to an increase in water level, with erosion at or below the water ~~line~~ by waves and currents can rapidly create an unstable bank (e.g. Terzaghi 1950).

Slope failures following rapid as well as repetitive drawdowns appear quite frequently in reservoirs (Terzaghi 1950, Jones et al. 1961, Erskine 1973). The effects of drawdown depend to a large degree upon the rate at



a. Initial ground water flow pattern and hydrostatic head h_i in colluvial cover over bed-rock.

b. Gradual filling alters pore pressures in the colluvium to approximately equal hydrostatic pressure (P) in the reservoir pool.

c. Drawdown reduces hydrostatic head of reservoir while hydraulic head of sediments responds with time by discharge at each water level (T_i). Rapid lowering in relation to low hydraulic conductivity results in much slower drop in water table.

Figure 28. Effects of water level fluctuation on ground water in shore zone materials (after Wahlstrom 1974).

which the water level drops and the permeability of the bank sediments (e.g. Wahlstrom 1974, Hopkins et al. 1975). Rapid reservoir drawdowns coupled with low permeabilities of bank sediments create particularly unstable situations, but slower drawdowns can also lead to failure conditions (Sherard et al. 1963, Zaruba and Mencl 1976, pp. 444-451).

If the water level is dropped slowly, the water table remains horizontal and descends at the same rate as the reservoir level. If water level is rapidly lowered, however, pore pressures drop more slowly than the hydrostatic pressure in the reservoir and this change results in a water table or piezometric surface that rises from the foot of the slope and is higher than that of the reservoir surface (Fig. 28). Flow of pore water would tend to be upward, outward and perpendicular to the piezometric surface in accordance with the hydraulic gradient i. Seepage pressures are generated by the water flowing from deeper to shallower bank sediments (Terzaghi 1950), and the resultant differential in pore pressure can also create an unstable condition.

In both cases, this reduction in shear strength or resistance to shearing is generally defined by the following equation (Terzaghi and Peck 1967):

$$s = c + (p - h\gamma_w) \tan\phi \qquad (6)$$

where s = the shearing resistance per unit of area
p = pressure per unit area at a given point on a potential surface of sliding due to the weight of the solids and water above that surface
h = the piezometric or hydraulic head at that point
γ_w = unit weight of water
tan ϕ = the internal angle of friction
c = cohesion.

If the material is mostly granular and therefore cohesionless, c becomes zero.

Thus, an increase in h reduces shearing resistance. If the excess hydrostatic pressure $h\gamma_w$ becomes equal to p, the total weight of the overburden is carried by the water, which cannot support it, and a failure takes place. Seepage pressures act in the direction of flow to reduce the effective stress.

In the case of reservoir drawdown, if the flow of pore water from the bank sediments does not keep pace with the water level drop and hydrostatic pressure decrease, unbalanced residual pressures develop in the bank sediments. Movement of the pore water under the hydraulic gradient generates seepage pressures which, combined with the residual pressures, can result in instability and failure along one or more failure planes. A similar situation can develop from a rapidly rising water table during heavy rainfall (Fig. 29). Movement due to failure on single or multiple planes may in turn cause vibrations and induce internal failure and collapse of the overlying sediments, perhaps leading to their liquefaction under certain conditions (Terzaghi 1950).

Ground water conditions, stability and failure modes are much more complex when bank and bluff sediments consist of interstratified cohesive and cohesionless materials with spatial variability in their dimensions and distribution. Then, failure of the entire bank can result from failure or liquefaction of a single sedimentary unit within it.

Terzaghi (1950) analyzed a slump-type failure condition along a single arcuate failure plane in fully saturated bank sediments (above and below the water table) that results from a rapid drawdown (illustrated in Fig. 30). All bank and bluff sediments are assumed to be fully saturated; this produces only a small error unless the sediments consist of coarse sand or gravel without admixtures of finer material. The potential failure surface

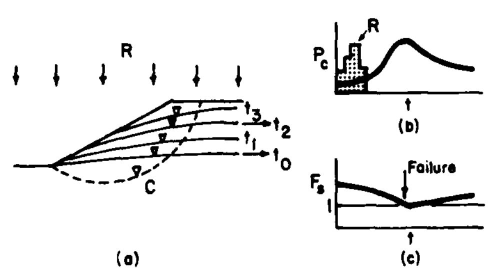
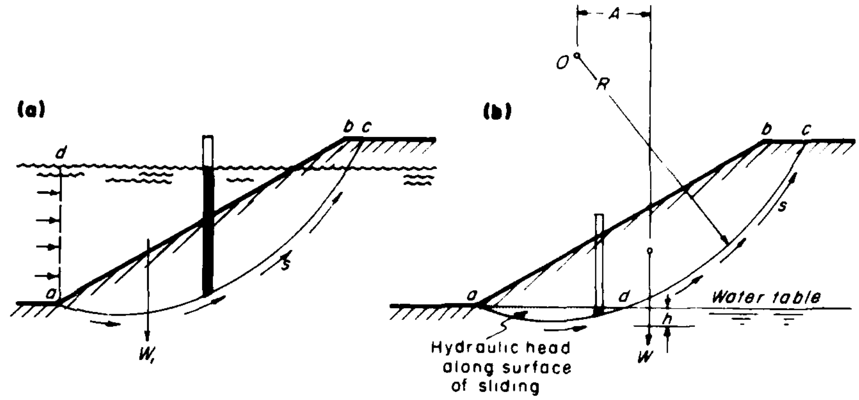


Figure 29. Changes in the position of the water table with time (after Freeze and Cherry 1979): a) changes due to heavy rainfall (R), b) pore pressures at point C during and following rainfall, c) the factor of safety (F_s) as a function of time.



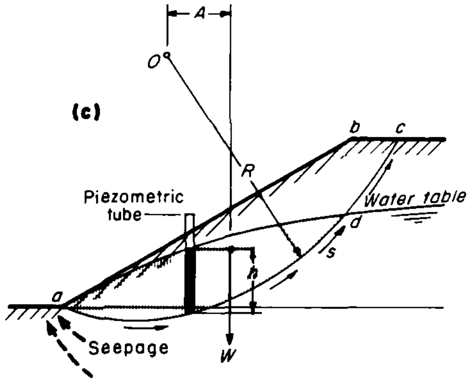


Figure 30. Analysis of slip-type failure of slope caused by drawdown. Initial (a), slow drawdown (b) and rapid drawdown (c) conditions are discussed in text (after Terzaghi 1950).

upon which sliding takes place is indicated by the arc ac and other variables are as follows:

- W = weight of slice abc, solid and water combined, per unit of length of the slope,
- ℓ = length of the arc ac,
- O = center of gravity of slice abc,
- c = cohesion of the slope-forming material,
- ϕ = angle of internal friction of this material,
- p = average unit pressure on the surface of sliding ac due to the weight W of the slice abc,
- h = piezometric head for any point of the potential surface of sliding at any time,
- h_1 = average of the piezometric heads h for the surface of sliding ac after a very slow drawdown
- h_2 = as for h_1 but after a rapid drawdown,
- A = resultant of all forces acting in horizontal direction, and
- R = resultant of all forces acting in direction shown in Fig. 30.

According to Terzaghi (1950),

> ...if the level of the body of water adjoining the slopes goes down very slowly, the water table remains horizontal and descends at the same rate as the water level of the reservoir. After the drawdown is complete, the piezometric surface is a horizontal surface

passing through the foot of the slope [Fig. 30b]. The average shearing resistance(s) of the material adjoining the surface of sliding is determined by eq [6], and the factor of safety of the slope with respect to sliding (G_s) is

$$G_s = \frac{R\ell\ [c + (p - h_1{}^{\gamma}w)\ \tan\phi]}{AW} \quad . \qquad (7)$$

On the other hand, if a drawdown takes place very rapidly the descent of the piezometric surface lags behind the descent of the free water level, and at the end of the drawdown the piezometric surface rises from the foot of the slope as indicated in Figure [30c] and intersects the potential surface of sliding at a point d which is located high above d in Figure [30b]. The corresponding factor of safety with respect to sliding G_s' is

$$G_s' = \frac{R\ell\ [c + (p - h_2{}^{\gamma}w)\ \tan\phi]}{AW} \quad . \qquad (8)$$

In Figures [30b and 30c] the total water pressure on the surface of sliding ac is indicated by shaded areas. Since the total water pressure on ac in Figure [30b] (slow drawdown) is very much smaller than that on ac in Figure [30c] (rapid drawdown), h_1 is very much smaller than h_2, and, as a consequence, G_s', eq [8], is smaller than G_s, eq [7]. Hence, even if a slope has survived a great number of slow drawdowns it may fail after a rapid drawdown, because G_s' is smaller than G_s.

As long as the piezometric surface in the ground beneath the slope has a gradient, the water percolates through the ground toward the surfaces adjoining the foot a of the slope, as indicated in Figure [30c] by dashed arrows. On account of its viscosity the percolating water exerts on the soil particles a pressure known as seepage pressure. This pressure acts in the direction of the flow, and its intensity increases in simple proportion to the seepage velocity. At the foot of the slope the seepage velocity and the corresponding seepage pressure are much greater than higher up, and the seepage pressure tends to move the soil particles along the flow lines which are directed toward the foot of the slope. As a consequence, at the foot of the slope, the point of failure is reached much earlier than at higher elevations, and once the lower part of the slope has failed, the upper part follows because it has lost its support.

In geologically complex sequences, fractures and other structural or sedimentologic features may locally complicate ground water conditions; further discussion of ground water-induced erosion and failure of bluff and bank zone slopes will be presented in later sections of this report.

Effects on vegetation

As the result of rising water levels or constantly fluctuating water levels, vegetation is killed by inundation (Fig. 27). This may occur during the winter (McKim et al. 1975), as well as during spring flooding. The vegetation cover and root system help dissipate wave, current and wind energy on low-lying beaches during storms and other temporary water surges. In addition, they resist subaerial erosion by overland flow during spring runoff and rainstorms, and their loss can result in accelerated erosion (e.g. Hunt et al. 1976, Pincus 1962, 1964). Nikolayenko (1974) observed that a dense shrub cover retarded or stopped breaking waves during storms on 3- to 4-m high bluffs. Furthermore, sediments in bluffs are effectively strengthened by extensive root systems and thus resist erosion and failure (Smith 1976). Berg and Collinson (1976) found that bluff recession actually lagged behind the rising water level in Lake Michigan in the late 60's because of protection by vegetation. With the onset of falling water levels, recession of the lake bluffs still continued until vegetation had reestablished itself.

Ice cover, frost and snow cover effects

Changes in water level have effects on the ice cover, snow cover and frost action on bluff and beach zone sediments. Measurements by Wolman (1959) and Hill (1973) indicated that a rise in river stage produced little erosion if banks were frozen, but if unfrozen and saturated, bank and bluff sediments (as, for example, exposed by reservoir drawdown) are more susceptible to erosion by currents and waves. Lowering the water level in winter also permits freeze-up, ice lens growth and heaving of the exposed sediments. Such particles loosened by frost action are highly susceptible to erosion by currents (Hill 1973). Thawing in the spring of ice-rich frozen sediments can also result in failure and flow of these thawed materials (e.g. Sterrett 1980, Reid 1984).

Rapid raising or lowering of the water level can also disrupt or even break up a complete ice cover (e.g. Billfalk 1982). A moderate decrease in water level can cause an ice cover to ground in shallow water areas and form nearshore cracks in the ice that approximately parallel the shoreline and bathymetric contours (e.g. Wuebben 1983a,b). These cracks may be perpetuated with recurring water level fluctuations. If the water that fills these cracks freezes, thermal expansion caused by temperature changes may force the ice cover against the beach or bluff, and bulldoze these sediments into unstable, erodable ridges. Similarly, a lowering followed by a rapid rise in water level would produce large blocks of floating ice that could be moved by the wind into the beach or bluff.

Lowering of the water to a stable level after ice cover formation drops the ice onto beach sediments, thereby protecting them from subaqueous and subaerial processes during the winter (Gatto 1982b), but possibly causing erosion from runoff during ice melt (Fig. 31). Any exposed sediments are susceptible to freezing and thawing, snowmelt runoff and other subaerial erosional processes in the spring. Conversely, raising the water after ice cover formation to a higher, stable level may cause ice to impinge directly on bluff sediments and also submit them to the effects of ice expansion and ride-up (Gatto 1982b).

Figure 31. Ice cover that has been fractured and partly broken apart by lowering onto beach zone sediments during reservoir drawdown in winter.

STABILITY AND MOVEMENT OF SLOPES

Rapid and sometimes extensive recession of bluffs may result from a loss of stability and the failure of bluff sediments (Fig. 32). Similar losses of materials below the waterline may also occur, often in concert with the failure and movement of overlying bluff sediments. Detailed observations of shore zone failure mechanisms and stability factors in reservoirs are limited to date, but some studies have analyzed the failure and movement of natural slopes and bluffs along rivers and lakes. The results of these studies are generally applicable to the problem of bluff stability and failure in reservoirs.

Figure 32. Rotational slip failure in bank material resulting in extensive bluff recession.

Stability

In general, every mass of sediment beneath a sloping ground surface or within vertically cut faces has the tendency to move downward under the influence of gravity. The resistance of the soil mass to the force of gravity determines the immediate stability of a slope; if it counteracts this force, the slope is considered stable. As defined by Casagrande (1936),

> Stability of a soil mass refers to the equilibrium of all external and internal forces with the resistance of the soil, including the force of gravity, seepage pressures, and any possible artificial disturbances due to construction activities, etc., as well as the effects of earthquakes. Stability does not refer to the amount of deformation which these forces produce, as long as the shearing resistance of the soil is not utilized to its ultimate limit.
>
> The stability of a mass of soil is not an individual property of the material like the specific gravity, permeability, compressibility or angle of internal friction, which can be measured on a sample of the soil and expressed by a single quantity. It is a combined effect of one or several of such individual properties and of numerous other factors, particularly the character of the forces to which the soil mass may be ex-

posed, its dimensions, various local conditions, and possibly other factors which are not sufficiently known ...

Thus the loss of stability or the creation of an unstable condition can result from processes that either modify the balance between the external and internal forces, or that alter the properties and hence shear strength of the materials themselves.

Because numerous and often complexly related factors can determine the stability of a particular slope, it remains extremely difficult to predict the stability or failure of natural slopes. In 1967, Peck concluded that the state of the art was such that it was impossible to reliably assess the stability of many, if not most, natural slopes under circumstances of practical importance. My review of the literature indicates that this situation has changed little to date. Thus, methods of stability analysis (e.g. Morgenstern and Sangrey 1978) require the careful and deliberate use of sound engineering and geological judgment in their application. Geologically complex situations remain largely beyond the scope of such analyses and a need still exists for basic scientific knowledge of the causes, mechanisms and critical, controlling factors of failures under natural conditions (Skempton and Hutchinson 1969, Patton and Deere 1971).

Loss of stability

The loss of stability and cause of slope movements along reservoir shores can result from 1) changes in material properties by degradational processes that reduce the strength of bluff or beach zone sediments, and from 2) external disturbances or erosional processes, such as the undercutting of the bluff, that reduce the resistance of the sediment mass to the force of gravity (e.g. Kachugin 1970). Water generally appears to be the most important agent altering material properties and reducing the shearing resistance of natural slopes (Cedergren 1977). The mechanics of bluff (or beach) failure depend upon the size, geometry and structure of the bluff (or beach) slope and the engineering properties of the bluff material (Thorne 1982). It is important to recognize that these factors are not static, but will change with time, possibly at varying rates as weathering and erosion take place (Vallejo 1977).

Along river banks, degradational and erosional processes are mainly fluvial entrainment at the bluff toe and interactive processes of weakening and weathering of the intact bank material (Thorne 1978, 1982, Hooke 1979, 1980). River bed degradation may also result in an effective oversteepening of an adjacent bank slope (Patrick et al. 1982). Along the Great Lakes' shores, similar bluff degradational processes are active (Edil and Vallejo 1977, 1980), but the principal erosional process at the bluff toe is wave action (e.g. Chieruzzi and Baker 1958, Hadley 1976, Hadley et al. 1977a,b, Vallejo 1977, Quigley et al. 1978, Edil and Vallejo 1980, Birkemeier 1981). Both situations provide analogs for bluff failures in reservoirs.

Reservoir bluff stability may be modified by undercutting and formation of a cavity or niche at its base by waves and nearshore wind-driven currents, especially during storms (e.g. Kachugin 1970, Reid 1984). Fluc-

tuating water levels allow the height of this niche to be greater than the wave height, assuming sufficient stability exists in the overlying sediments to allow for niche growth. Generally sands or other cohesionless materials tend to slough from a face almost immediately (e.g. Fisk 1952, cover photo). Gradual reduction in stability of cohesive sediments by toe erosion may result in failure by slip or flow (e.g. Hutchinson 1983). Cohesive sediments may resist failure sufficiently to allow blocks of material to be cantilevered over the water surface until some critical distance is reached at which the overlying material fails and the block falls (e.g. Thorne 1978, Birkemeier 1981). Joints, fissures or other near-vertical discontinuities may act as planes of weakness along which failure may take place before the material itself losses its stability (e.g. Deere 1957).

Sheet flow, rill erosion and gullying are important in physically removing sediment from the bluff face and can result in the loss of stability in the remaining particles or blocks of sediment between the rills or gullies. These processes may also affect soil moisture and groundwater flow conditions (e.g. Vallejo 1977). Raindrop impact and creep have similar effects. Wind may also winnow out finer sediments when bluff materials are dry, thereby freeing coarser particles to fall to the base of the bluff. In composite banks, layers of cohesionless materials may selectively fail, causing each overlying cohesive layer to be cantilevered over cavities formerly filled with the cohesionless materials. Locally, removal of bluff sediments by erosional processes reduces overburden pressures and may allow stress relief and spalling in fissured or jointed consolidated sediments. For each of these processes, their overall impact will depend upon factors such as slope angle, extent of the vegetation cover, soil moisture content, material types and properties, and local climate (e.g. Carson 1971, Bodenko et al. 1978, Thorne 1980, Edil and Haas 1980).

Internal changes in the strength of bluff sediments can result from several processes not always evident at the bluff surface. Of primary importance are changes in soil moisture, pore pressures and seepage pressures. Seepage, in general, reduces slope stability. Zaslavsky and Sinai (1981) have concluded that seepage may actually predominate in causing surface erosion, including that within rills. These changes can be brought about by a rise in the water table, as when the water level rises in the adjacent impoundment, or during extended periods of higher precipitation (e.g. Kachugin 1970, Carson and Kirkby 1972).

Movement of water through bluff materials can leach out soluble chemicals or clay particles, thereby reducing their shear strength with time (e.g. Terzaghi and Peck 1967, Kachugin 1970). Leaching may be especially important in loss of stability of sensitive or quick clays (e.g. Smalley 1976, Carson 1977).

Similarly, seepage pressures due to ground water flow out of bluff faces may reduce their effective stress. Piping can wash out near-surface sediments within the bluff face, thereby undermining other material within the face (e.g. Deere 1957, Kachugin 1970, Edil and Vallejo 1980). Reduction in shear strength sufficient to cause failure may be confined within fracture zones or individual strata as pore water pressure rises and may result in slip along or within such effective planes or zones of weakness (e.g. Casagrande 1936, Terzaghi 1950).

Freezing of water within pores or fissures heaves soil particles apart, physically loosening or detaching the sediments and reducing strength derived from both particle interlocking and cohesion (Wolman 1959, Corte 1969). Thawing of ice-rich sediments can produce saturated or over-saturated materials that are near failure, with excess pore pressures generated just above the thawing interface (e.g. McRoberts and Morgenstern 1974a). Simple alternating wetting and drying may also cause the loosening and slaking of exposed bluff materials. Bluffs composed of interstratified cohesive and cohesionless sediments exhibit much more complex behavior because individual layers may be more susceptible to internal changes in strength than the sediment mass as a whole.

Types of movement

The types of failure will obviously vary with the configuration and physical properties of the bluff as well as the forces tending to cause failure. Slope movements range from those in which single particles, aggregates or blocks of material undergo failure and free fall, to those involving the en masse flow of saturated, remolded material (Fig. 33, cover). The mechanisms involved in the movements are often complex. Because of the apparent continuum between many of the slope processes, distinguishing individual types of movement is somewhat difficult.

Several classifications separating failure modes exist; no matter which is chosen, they all are somewhat arbitrary because of the gradational nature of the processes and the variability inherent in materials composing slopes. A widely accepted classification proposed by Varnes (1958, 1978) provides a useful format for describing the basic types of movement. Oth-

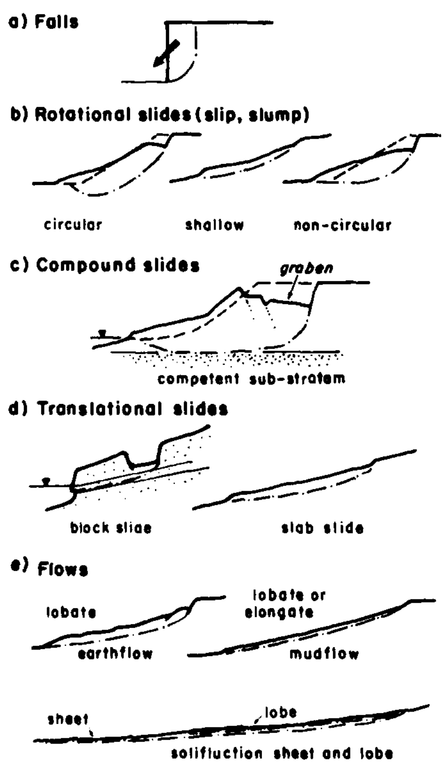


Figure 33. Cross-sectional profiles of some basic types of slope failures as defined by Skempton and Hutchinson (1969) for clay slopes. Failure surface shown by dashed and dotted line.

ers, such as those of Sharpe (1938) and Skempton and Hutchinson (1969), are similar in concept.

Under the Varnes scheme, the primary types of slope movement include falls, topples, slides, lateral spreads and flows, and complex types involving two or more of these (e.g. slides leading to flows). Briefly these movements are as follows:

1. Falls - A material mass is detached from a steep slope or cliff and descends primarily through the air to the slope's base by free fall, leaping, bounding or rolling.

2. Topples - Large blocks or segments of the slope undergo a forward rotation about a pivot point, following an upslope tensional fracturing and failure due to the force of gravity, caused by forces exerted by adjacent sediments or by a wedging force in cracks. Forward rotation culminates with a fall or perhaps slide to the slope's base.

3. Slides - Sediments move downslope under the force of gravity by slip along one or several discrete surfaces or within a thin zone at the base of the moving material. Two types are recognized. Rotational slides or slumps rotate down and out along a surface that is roughly concave upward. Translational slides move down a more or less planar surface such as defined by joints or bedding planes, without a rotary motion. These types are also called slab slides. Transitional forms between these end-members are common, with sliding surfaces often controlled by geologic discontinuities.

4. Lateral spreads - The dominant mode of movement involves a lateral extension of the slope sediments accommodated at the upslope end by a shear or tensile fracture. Fine-grained sediments, particularly sensitive silt or clay that loses most or all shear strength on disturbance or remolding, exhibit lateral spreading, usually as a progressive failure. Thus a slump along a shore initiates progressive failure that extends retrogressively landward into the bluff. Lateral spreads range gradationally between block slides and flows.

5. Flows - In a general sense, flows are sediment/water mixtures that move downslope under the force of gravity. They appear to be continuously deforming and without distinct slip surfaces. Rates of movement may range from almost imperceptible to extremely rapid. Flow characteristics, including rate, style and form, can vary continuously and

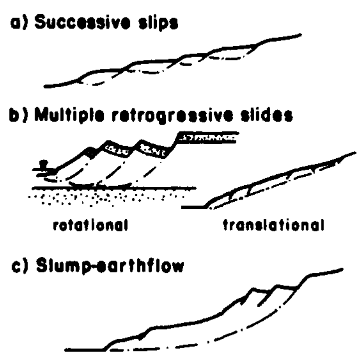


Figure 34. Three examples of common, complex slope movements in cross section.

appear related to the water content. They may occur subaerially or subaqueously, but exhibit different mechanisms and form.

6. Complex - Complex flows are slope movements involving two or more of the principal types of movement listed above, leading to one another during the course of movement, or simultaneously occurring within different parts of the same moving mass (Fig. 34). Slides leading to lateral spreads or flows are fairly common types described in the literature.

Mechanics

The mechanics of each type of movement are complex and remain poorly understood, particularly in regard to flows, lateral spreads and progressive or retrogressive slides and flows. Certain slope movements with simplified boundary conditions can, however, be treated mathematically by the technique of limit equilibrium (e.g. Terzaghi and Peck 1948), providing some insight into failure conditions.

As an example, Thorne (1978, 1982) and Thorne and Tovey (1981) have analyzed the mechanics of failure for slope movements that occur along actively eroding riverbanks, both those being undercut by currents and those removed from the immediate effects of currents, using fundamental soil mechanics theory and the concept of effective stress.

Basic failure conditions. Thorne (1978, 1982) considers three classes of riverbanks based upon their general composition: cohesive, cohesionless or composite (consisting of interbedded cohesive and cohesionless sediments). In each case, stability is assumed to depend upon a balance between motive and resistive forces associated with the most critical mechanism of failure, which can vary with the size, geometry, structure and engineering properties of the bank, external forces and climatic conditions. Vallejo (1977) applied slope stability theory in a similar manner to analyze bluff stability and failure on Lake Michigan.

For both cohesive and non-cohesive materials, the shear strength is given by the revised Coulomb equation in terms of effective stress (Terzaghi and Peck 1967):

$$s = (\sigma - u) \tan \bar{\phi} + \bar{c} \qquad (9)$$

where s = shear strength
u = pore pressure
σ = normal stress on the shear surface
$\bar{\phi}$ = effective angle of internal friction
$\bar{c}$ = cohesion intercept.

In the case of noncohesive materials, $c = 0$ and drops out of the equation. Shear strength parameters are usually defined through testing.

For noncohesive materials under drained conditions, the pore water pressure is negligible and $u = 0$. Stability of a slope of an infinite length then depends upon the angle of slope and angle of internal friction. A factor of safety F_s can be defined by

Figure 35. Bluff consisting of mostly noncohesive sand and gravel undergoing failure by individual grain falls and shallow, localized slips. Material accumulates at base as talus deposit, which protects the slope from further undercutting and failure until removed by waves and currents.

$$F_s = \frac{\tan \theta}{\tan \phi} \tag{10}$$

where θ is the slope angle. The limiting case occurs where $\phi = \theta$ and the slope is at the point of failure. F_s greater than one indicates stability.

Failure of drained noncohesive banks takes place when either the friction angle is reduced to less than the slope angle, or the materials are effectively oversteepened so that $\theta > \phi$. The latter case exemplifies the undercutting of bluff materials by currents or waves, while the former reduction in ϕ can result from weakening and weathering of the slope-forming sediments (e.g. Deere and Patton 1971). These processes reduce the packing density and hence friction angle of the material. Most failures in homogeneous, noncohesive banks involve the fall of individual grains, or result from shallow slips along planar surfaces because the shear strength of these materials will usually increase with depth faster than shear stress (Thorne 1982) (Fig. 35).

Under undrained conditions, pore water pressures may be important in determining shear strength. The limiting slope angle is then given by

$$\tan \alpha = \frac{(\gamma z_p \cos^2\theta - u)\tan \bar{\phi}}{\gamma z_p \cos^2\theta} \qquad (11)$$

where γ is bulk unit weight of bank material, and z_p vertical depth to the failure plane (Carson and Kirkby 1972). Thus where positive pore pressures are present, the limiting slope angle will be less than the angle of internal friction. Such positive pore pressures can develop, for example, when submerged materials are uncovered by rapid drawdown. Under partially saturated conditions, the pore water pressure term is negative and imparts an apparent cohesion due to capillary effects (Lambe and Whitman 1969). For sand or coarser well-sorted sediments, this apparent cohesion can be considered negligible.

The effect of flowing ground water upon slope stability can often be approximated by assuming that flow is approximately parallel to the slope. Where appropriate data can be measured, however, construction of a flow net based upon the pore water pressures is a more accurate representation and clearly needed in slopes with geological complexities (Cedergren 1977). Calculations may also be performed based upon an analysis of the forces acting on an element within an infinite slope (i.e. thickness of unstable, failing material is small compared to the height of the slope). For forces in equilibrium

$$\frac{T}{\bar{N}} = \tan \bar{\phi} = \frac{\gamma}{\gamma_b} \tan i \qquad (12)$$

where $\bar{\phi}$ = friction angle based on effective stress
T = shear force
$\bar{N}$ = effective normal force
γ = total unit weight
γ_b = buoyant unit weight
i = seepage gradient.

Since γ_b/γ of clean sand is about 1/2, the maximum possible stable slope under these conditions is about 1/2 $\bar{\phi}$. Thus in an infinite slope composed of sand with the above conditions, seepage reduces the maximum stable slope angle to about half that for sand that is dry or is completely submerged and without ground water flow (Lambe and Whitman 1969).

Thorne (1978) observed failures that were similar to those of non-cohesive, drained banks but that resulted from increases in pore water pressure such as at the base of high banks or slopes after a heavy rain, during snowmelt and runoff, or after a rapid drawdown in the water level. Shallow slips and individual grain failures were common. He also commented on the importance of high seepage pressures, which cause piping by physically removing the sediment at the bank face where water emerges (described as "sapping" by Sterrett [1980] and others), or by creating excess hydrostatic pressures at locations where water percolates upward at the base of a slope and exceeds the effective weight of the overlying sediments. The mechanics of piping will be described in Ground Water Erosion.

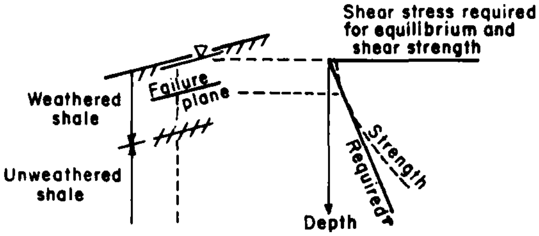


Figure 36. Failure plane development within weathered soil or rock (after Lambe and Whitman 1969). Shear stress and shear strength changes with depth are indicated.

Stability of cohesive bluff sediments strongly depends upon the height and slope angle of the bluff and the strength parameters at various depths (Lambe and Whitman 1969). Solving eq 12 for an infinite slope with appropriate values for s and (σ - u) required for equilibrium,

$$\frac{\bar{c}}{\gamma\ H_c} = \cos^2 i\left(\tan i - \frac{\gamma_b}{\gamma}\tan\bar{\phi}\right) \tag{13}$$

where H_c is depth to the failure plane. Under submerged conditions without seepage $\gamma = \gamma_b$ and

$$\frac{\bar{c}}{\gamma_b\ H_c} = \cos^2 i\left(\tan i - \tan\bar{\phi}\right)\ . \tag{14}$$

Equations 13 and 14 have been applied to approximate failure conditions in overconsolidated cohesive sediments at relatively shallow depths and in deeper sediments where cohesion and strength are lower than in overlying materials. Such a situation may exist in nature on long slopes weakened at depth by weathering or where soil moisture increases rapidly during heavy precipitation (Fig. 36). It is also applicable to sediments lying directly on rock at a shallow depth (Lambe and Whitman 1969).

For failure conditions where H_c approaches the actual height of the slope in cohesive materials, the infinite slope is no longer appropriate and the problem must be treated as one of limited height. This involves use of Bishop's (1955) method of slices (see Rotational and Plane Slip Failures below).

Thorne (1978, 1982) has suggested that a Culmann analysis can be used for plane slip failures of banks that are steep to vertical and low in height (Fig. 37). For the Culmann analysis, the failure plane is assumed to be planar and pass through the toe of the slope (Fig. 38). Resolving the forces along and normal to this failure plane allows calculation of the critical bank height H_c:

$$H_c = \frac{4c\sin\theta\cos\phi}{\gamma\ [1 - \cos\ (\theta - \phi)]}\ . \tag{15}$$

This equation simplifies to

$$H_c = \frac{4c}{\gamma}\tan\ (45 + \phi/2) \tag{16}$$

for banks (bluffs) with vertical faces.

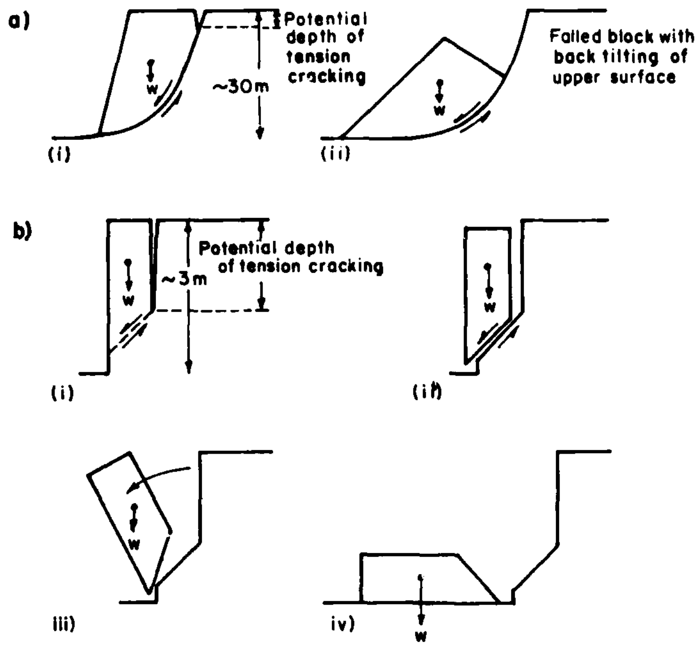


Figure 37. Progression of rotational slip failure in high bank with steep face (a), and plane slip or toppling failure (b) in bluff that is low in height with steep to vertical face (after Thorne and Tovey 1981).

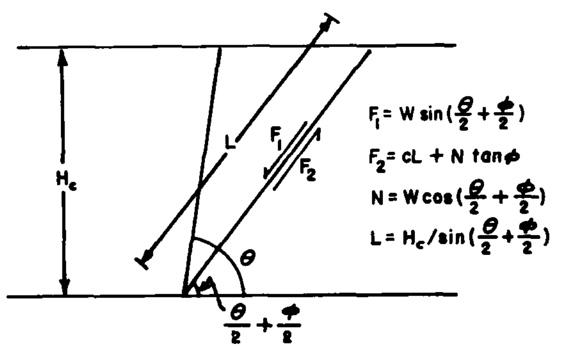


Figure 38. Culmann analysis for plane slip failure in banks of low height with near-vertical face (after Thorne 1982). Parameters defined in text.

This technique, in contrast to others discussed here, uses the total stress, rather than effective stress, for calculating stability and is generally not recommended for use in analyzing long-term slope stability (Lambe and Whitman 1969). It does not consider the effects of pore water pressure variations directly, but field studies by Bradford and Piest (1977), Slavin (1977) and Thorne (1978) have suggested that pore pressures are probably not an important factor in causing cantilever failures, at least in the case of low river banks subject to undercutting.

Thorne (1978, 1982) has also suggested that a modified Culmann analysis as first described by Terzaghi (1943) can be applied to the case of

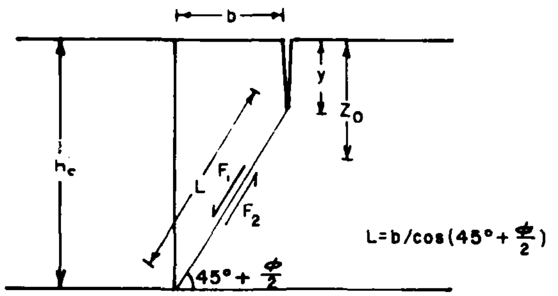


Figure 39. Modified Culmann analysis for plane slip failures in banks of low height, with near-vertical faces that are affected by tension cracking (after Thorne 1982). Parameters defined in text.

tension cracking in vertical bluffs (Fig. 39). Slope stability is reduced by cracking because of a decrease in available cohesion along the upper part of a potential failure surface (Lambe and Whitman 1969). The depth to which sediments are affected by tensile stress (z_o) is given by the equation

$$z_o = \frac{2c}{\gamma} \tan (45 + \phi/2) \tag{17}$$

and empirical evidence suggests that this maximum depth of cracking is about half the height of the vertical face (Terzaghi 1943). The result of such tension cracking in low vertical banks is for the outer slab of soil to become detached and then slip downward and outward on a plane or slightly curved failure surface (Thorne 1978). The critical height for a tension-cracked vertical bank (H_{CT}), assuming minimal tensile strength, is

$$H_{CT} = H_c - z_o = \frac{2c}{\gamma} \tan (45 + \phi/2) \ . \tag{18}$$

But since tensile stress is at a maximum at the ground surface and decreases linearly with depth to zero at z_o, the actual depth of cracking (y) for soil of a tensile strength of σ_{TC} is (Lohnes and Handy 1968)

$$y = Z_o \ (1 - \sigma_{TC}/\sigma_T) \tag{19}$$

where σ_T = tensile stress at surface. Since

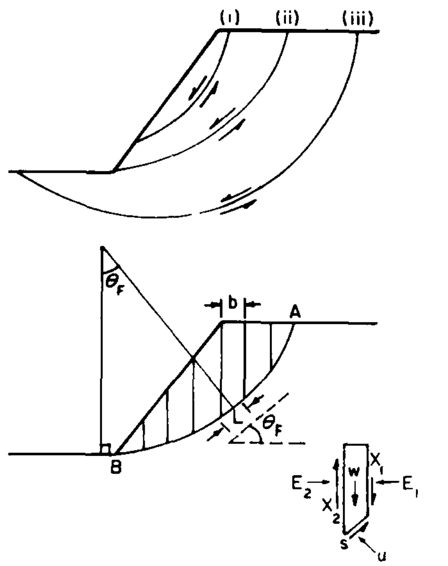
$$\sigma_T = 2c \tan (45 - \phi/2) \tag{20}$$

then the width of the failure slab b (Fig. 39) is given by

$$b = \left(\frac{H_{CT} - y}{\tan (45 + \phi/2)}\right) \ . \tag{21}$$

This calculation is only approximate, however, because once cracking begins, the stress distribution is altered and continues to change as the depth of cracking increases (Lohnes and Handy 1968).

Field application of this analysis by Thorne et al. (1981) indicated that tension crack depth and stability were reasonably well predicted by these equations using field data on bank height, slope angle and soil properties from streams in northwestern Mississippi.



a. Failure arcs indicated for (i) thin slope failure, (ii) toe failure and (iii) base failure.

b. Stability analysis of a slip circle by the method of slices with restoring and disturbing forces resolved for a single slice i shown on the lower right. X represents internal forces and E external forces.

Figure 40. Rotational slip failures in cohesive slope materials (after Thorne 1982).

Rotational and plane slip failures. The classic rotational slide on a circular failure surface has been analyzed in a number of ways, the commonly used methods often being the conventional method of slices described by Taylor (1948) and Bishop's simplified method of slices (Bishop 1955). The method of slices is applicable for analyzing long-term stability of natural slopes where failure eventually results from longer-term changes in strength parameters or pore water pressure and where the critical height H_c approaches the actual height of the slope (Fig. 37a). Effective strength parameters are therefore used in these analyses so that measurements of pore water pressures are needed.

Bishop's (1955) method of slices appears accurate for most purposes and is commonly used (Morgenstern and Sangrey 1978). For the method of slices, the normal stress acting at a point on the failure arc is assumed to be determined mainly by the weight of soil overlying it. For a particular slope, the mass of material above the circular failure plane is divided into a series of vertical slices (Fig. 40) and the equilibrium of each slice, based upon the forces acting on it, is defined. Because of the number of unknown factors involved, simplifying assumptions are made. For Bishop's method, the forces acting on the sides of any slice are assumed to act horizontally.

For this method, a safety factor F is defined as a ratio of restoring moments to disturbing moments about the center of the failure arc:

$$F = \frac{\text{Moment of shear strength along failure arc}}{\text{Moment of weight of failure mass}} = \frac{M_R}{M_D} \quad . \qquad (22)$$

The factor of safety for Bishop's method is given by

$$F = \sum_{i=1}^{i=n} \frac{\left[\bar{c}\ \Delta x_i + (w_i - u_i\ \Delta x_i)\ \tan\bar{\phi}\right]\left[1/M_i(\theta)\right]}{\sum_{i=1}^{i=n} w_i \sin\theta_i} \tag{23}$$

where

$$M_i\ (\theta) = \cos\theta_i\quad 1 + \frac{\tan\theta_i \tan\bar{\phi}}{F} \tag{24}$$

and u_i is pore water pressure at the base of slice i. F is solved in an iterative manner, but the values converge rapidly. Several possible locations of the critical slip circle must be tried because there is no way to define its location based upon these relationships. Bishop and Morgenstern (1960) and Morgenstern (1963) developed stability charts, the latter for undrained conditions, to predict the worst case.

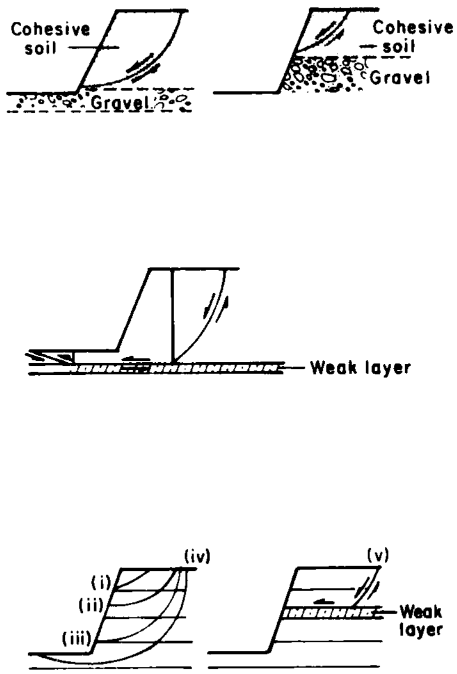
In slopes where noncircular slip surfaces of irregular shape and unknown geometry are anticipated, such as in bluffs of layered cohesive sediments, of heavily fissured materials or with individual layers or zones of low strength, techniques such as defined by Morgenstern and Price (1965) are more appropriate. Morgenstern and Price's technique also employs the method of slices but accounts for all boundary and equilibrium conditions. To make the analysis statistically determinate, the shear and normal forces X and E acting on each slice are assumed to be related by the expression

$$X = \lambda\ f(x)\ E \tag{25}$$

where λ is a scale factor determined by the solution and $f(x)$ is an arbitrary function concerning the distribution of the internal forces. For each solution, the choice of $f(x)$ is limited by conditions of physical admissibility, which require that no tension be developed in the sediments above the failure plane and that the failure criterion as defined not be violated. To solve these equations requires extensive iterative calculations best suited for a digital computer. The reader is referred to the original reference and Morgenstern and Price (1967) for a discussion of the numerical method for solving the equations.

A more recent approach, as yet requiring field verification of its applicability, was presented by Baker and Garber (1978). They analyzed the stability of slopes on the basis of the concept of the limiting equilibrium state but using variational calculus to derive a general theorem governing the shape of potential slip surfaces. This theorem is valid for the general case of nonhomogeneous and nonisotropic soils with an arbitrary distribution of pore water pressure and external loads. Baker and Garber selected this technique because it involved no a priori assumptions on either the shape of the slip surface or the stress distribution on it.

Baker and Garber (1978) concluded that two potential families of slip surfaces exist for the two failure modes of translational or rotational slides, and that the shape of the slip surface is governed by two differential equations. Because each has special or unique geometrical properties, the directions of elementary forces are defined in each case. They apply



a. Rotational slip failures, with slip at gravel bed interface.

b. Composite slip surfaces due to weak layer.

c. Multiple possible failure planes within multilayered bluff sequence of differing strengths and containing a single weaker layer that overrides the influence of other possible failure planes.

Figure 41. Examples of possible rotational slip failures in high composite bluffs (after Thorne 1982). Bedding planes, weak horizons and other discontinuities may act as failure planes.

to sediments with an arbitrary distribution of c, ϕ, γ, μ, slope and external loads; only the distribution of ϕ affected the nature of the potential slip surfaces and other properties determined which failure surface was the critical one. Baker and Garber (1978) also defined a method for determining the factor of safety that is less laborious than the previous method of Morgenstern and Price (1965) and similar to that of Bishop (1955).

The stability of bluffs composed of interstratified cohesive and cohesionless sediments is often much more difficult to analyze. Geologic controls on where failure may take place are exceedingly important and thus the location and shape of the failure plane are much harder to define. While many riverbanks exhibit several repetitive sequences of alluvial sediments (e.g. Fisk 1952, Brunsden and Kesel 1973, Thorne and Tovey 1981, Turnbull et al. 1966), sediments composing shore zones of reservoirs in northern regions can also consist of complex glacial, periglacial, colluvial, and lacustrine deposits. Their geologic history may influence and complicate stability calculations for eroding bluff faces (e.g. Deere and Peck 1958). Sterrett and Edil (1982) have discussed the application of Bishop's method (1955) and its problems to bluffs composed of glacial deposits with complex ground water flow systems.

As stated previously, rotational slip failures along noncircular surfaces often occur because of various geologic discontinuities in composite bluffs or banks (Fig. 41). Static analyses discussed above for the cohesive or cohesionless cases apply to individual strata, but the problem of analyzing composite bluffs requires identification of where the actual

failure plane is likely to be located. A trial-and-error technique must be applied to composite slopes because the critical failure surface may occur within one layer or between several layers of multilayered (stratified) bluffs, as well as for the simpler case where a single weak layer lies within the slope but the point of intersection of the failure plane and this layer are unknown. Noncircular failure surfaces may be analyzed by the methods of Morgenstern and Price (1965), Sarma (1979), or Baker and Garber (1978), but field experience and testing are still needed to determine the accuracy of each method for natural slopes.

Plane slip failures are apparently common in slopes of high river bluffs with multiple thin cohesive layers, low river bluffs in general (Thorne et al. 1981, Thorne 1982) and low angle slopes with well-developed clay-rich, B horizons (Trimble 1979). Similar conditions were described for bluffs bounding the Great Lakes (e.g. Chieruzzi and Baker 1958, Carter 1976, Birkemeier 1981).

Fall and topple failures. Thorne (1978) and Thorne and Tovey (1981) discuss the failure mechanisms of vertical slopes developed in cohesive sediments which are underlain by cohesionless sediments. Erosion of these underlying cohesionless sediments can generate an overhanging or cantilevered block of the cohesive material (Fig. 42).

Three principal modes of failure were recognized as resulting from continued expansion of the niche or cavity or weakening of the overlying

Figure 42. Eroding bluff exposed by lowered wate[illegible] reservoir. Cohesive upper unit overlies niche or cavit[illegible] o[illegible] 0.6-m height that is developing at its base (darker area, low[illegible] t).

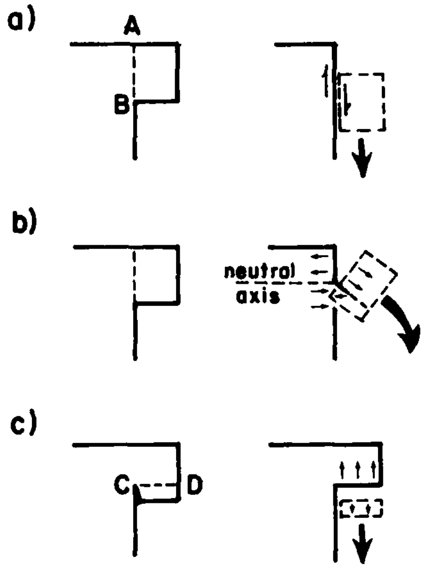


Figure 43. Principal modes of failure of cohesive sediments that are cantilevered by erosion of underlying cohesionless sediment (after Thorne and Lewin 1979). Shear (a), beam (b) and tensile (c) failures.

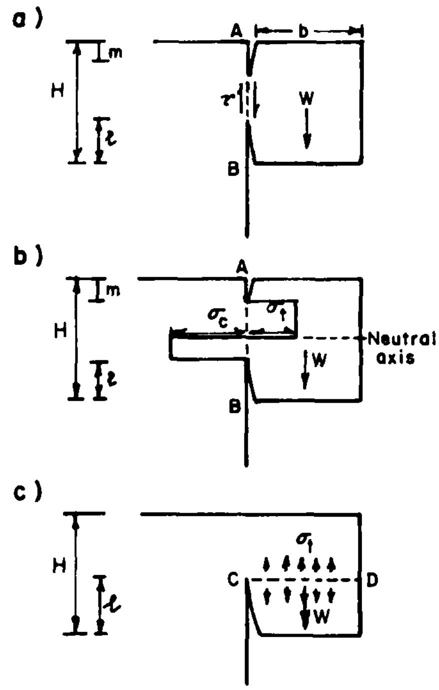


Figure 44. Forces of weight, shear, compression and tension acting on cantilevered sediments for (a) shear failure, (b) beam failure and (c) tensile failure. Parameters defined in text (after Thorne and Tovey 1981).

sediments by wetting or cracking (Fig. 43). A shear failure can occur by downward displacement of an overhanging block along a vertical plane AB and result from the shear stress due to weight of the block overcoming the shear strength of the sediment along that plane (Fig. 43a). In a beam failure (Fig. 43b), a block rotates forward about a horizontal axis in the block. At this axis, forces are neutral while above the axis, the block is in tension and below it in compression. Once the moment of the weight of the block about the neutral axis overcomes the resistive moments of the soil's strength in tension and in compression, failure takes place. In the final case (Fig. 43c), a tensile failure occurs across a horizontal plane within the overhanging block and it falls when the tensile stress of the failed block overcomes the tensile strength of the sediment.

Beam failures were the types of failure most commonly observed by Thorne and Tovey (1981). Shear failures generally occurred in cohesionless sands or where root systems in bank vegetation provided little strength to the overhanging block. The tensile failures commonly took place after vertical cracking up from the cavity along vertical planes of weakness within the cantilevered material, followed by failure along horizontal planes of weakness or zones of minimum tensile strength. Desiccation after exposure of the sediment to the air was also apparently important.

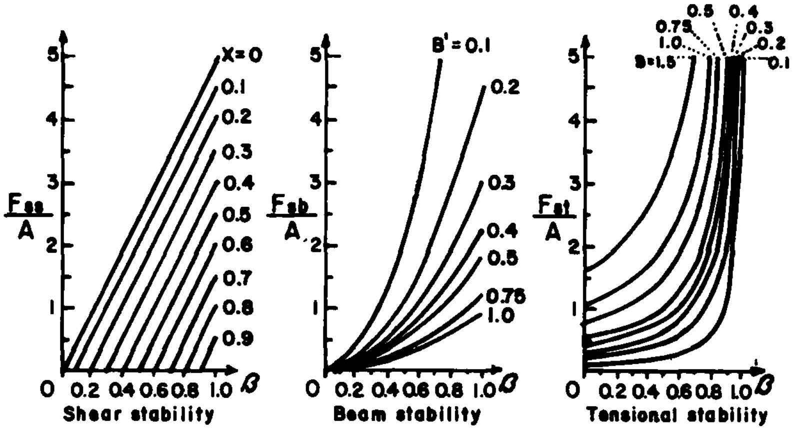


Figure 45. Dimensionless charts for cantilever stability. Parameters defined in the text (after Thorne and Tovey 1981).

Thorne and Tovey (1981) discussed cantilever stability based upon a static equilibrium analysis of forces and the theory of bending beams (Fig. 44). Forces of weight, shear strength, compressive strength and tensile strength were resolved horizontally and vertically, and moments about the neutral axis were determined for shear, beam and tensile failures of the block. Dimensionless charts for cantilever stability (Fig. 45) were derived from the following stability equations (Thorne 1978).

For shear failure:

$$\frac{F_{ss}}{A} = \frac{(\beta - \chi)}{\beta} \frac{\beta}{2r} \quad . \tag{26}$$

For beam failure:

$$\frac{F_{sb}}{A} = \frac{\beta^2}{(1 + r)\, B'} \quad . \tag{27}$$

For tensile failure:

$$\frac{F_{st}}{A} = \frac{B}{(1 - \beta)} \quad . \tag{28}$$

where

$$A = \frac{\sigma_t}{\gamma_b} \tag{29}$$

and r is the ratio of tensile strength to compressive strength, and F_{ss}, F_{sb} and F_{st} are factors of safety for shear, beam and tensile failure respectively.

The dimensionless numbers β, χ, B and B' depend upon cantilever geometry as follows:

$$B = \frac{b}{H} \quad (30)$$

$$B' = B \left[\frac{\beta}{\beta - \chi}\right]^2 \quad (31)$$

$$\beta = \frac{H - \ell}{H} \quad (32)$$

$$\chi = \frac{m}{H} \quad (33)$$

where H is overhang (cavity) height, m is length of the upper crack and ℓ is length of the lower desiccation crack (Fig. 44). The stability charts (Fig. 45) are based upon an r-value of 0.1, but Thorne and Tovey (1981) indicated that block stability is rather insensitive to changes in r.

The stability charts allow identification of the most critical mechanism of failure and the limiting factor of safety. The values of A, B, β' and χ parameters must be calculated if the charts are used, and the geometry of the cantilever block and its soil properties must be measured. Thorne and Tovey (1981) present examples of such calculations for banks on the River Severn in Wales. Test banks suggested that prediction of the factors of safety were within 10 to 15% of the true value and that the most critical failure was correctly assessed.

Lateral spreads and retrogressive failures. Laterally spreading slope movements typically form in fine-grained sediments on shallow slopes and

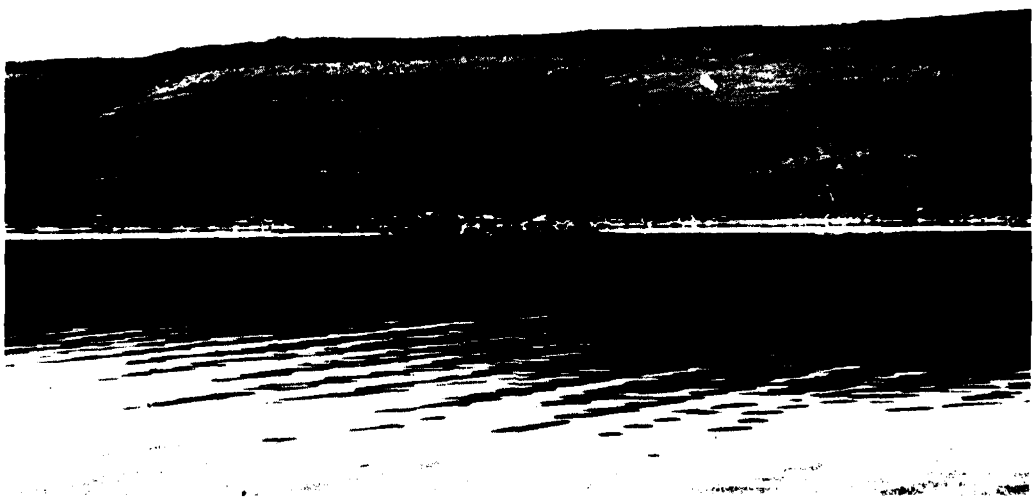


Figure 46. Progressive slump-flow failure of low bluff and adjacent, landward sediments on Lake Ashtabula, N.D., apparently initiated by toe erosion at higher water level.

occur rather rapidly with little warning (Varnes 1978). Sensitive silt and clay quickly lose shear strength upon disturbance and remolding and are unique types of materials characterized by spreads (e.g. Bjerrum 1955, Crawford and Eden 1967, Cabrera and Smalley 1973, Mitchell and Markell 1974, Mitchell and Klugman 1979). Such silts and clays have properties that cause them to be particularly sensitive, although controversy remains as to which properties are more important and how they actually alter or produce sensitive materials (e.g. Smalley 1976, Kerr 1979, Gillott, 1979, Moon 1979).

Failures may take place gradually over tens of years and are progressive, starting at one location and spreading laterally into previously undisturbed sediments (Fig. 46). The initial failure may involve a distinct rotation, but quite often the principal movement is one of translation (e.g. Thomson and Hayley 1975, Haug et al. 1977, Bjerrum 1971, Carson 1977). Many of the reported progressive failures start initially as a slip failure of a streambank or lake bluff, and then spread landward into undisturbed sediments. Progressive or retrogressive failures commonly move along a noncircular failure surface along which its peak or maximum shear strength has been reduced by large strains applied during previous downslope failures to its residual strength, with each successive failure following that same surface (Fig. 47) (Bjerrum 1967, 1971, Bishop 1967, Carson 1977). Many of the reported progressive failures start initially as a slip provide lateral support to upslope blocks (Fig. 48). In others, failure and remolding are rapid so that most material loses its structure and flows almost immediately (Skempton and Hutchinson 1969).

The mechanisms of these retrogressive failures are poorly understood. Bjerrum (1967, 1971) concluded that such failures in overconsolidated clays are preceded by the development of a continuous sliding surface caused by a mechanism of progressive failure which reduces the undrained shear strength to its residual value. Analyses of progressive or retrogressive failures have indicated that the average shear stress causing failure is smaller

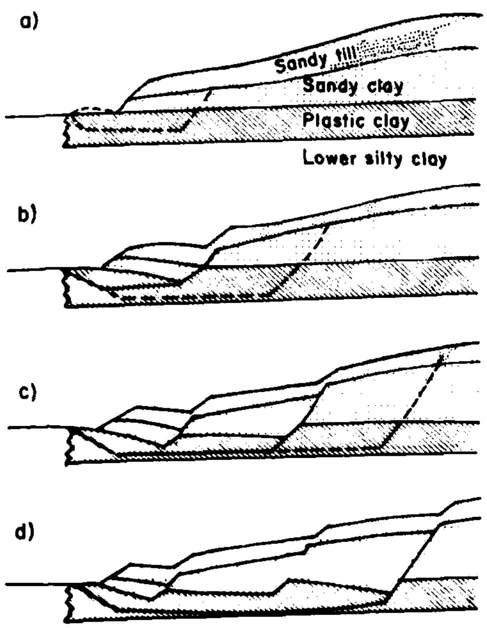


Figure 47. Sketch of the progressive slip failure of stratified unconsolidated glacial deposits overlying clay and silty clay deposits on a noncircular failure surface that was initiated by excavation of the lower part of the slope (after Bjerrum 1967). Each successive failure moved along the same lower surface; the actual sequence developed gradually over 80 years and consisted of many more such slip failures.

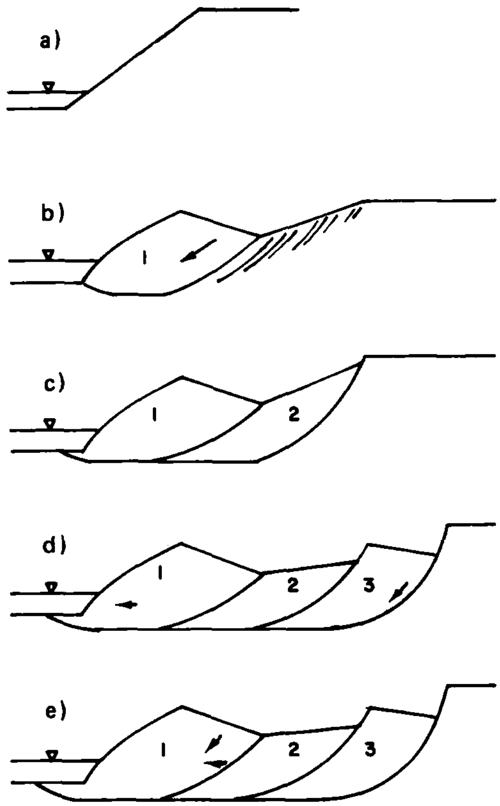


Figure 48. Example of retrogressive slope failure due to toe erosion along a river as determined by Haug et al. (1977): a) undisturbed slope, b) rotational failure and pulling down of the upslope scarp, c) rotational failure of block 2, d) horizontal movement of block 1, allowing block 2 to settle and block 3 to fail, and e) additional rotation and translation of block 1 with blocks 2 and 3 further settling.

than the shear strength of the failed material, but it is typically similar in magnitude to the residual stress of that material.

Conditions apparently required for the development of the continuous failure surface with a residual strength are 1) an internal discontinuity or external disturbance where failure can first take place and 2) material properties and behavior such that a) local internal shear stresses must exceed the peak shear strength of the material, b) local differential strain must exceed the strain at which the material will fail for the failure surface to advance, and c) a rapid and large decrease in shear strength must result with strain after failure, so that shear resistance in the failed zone does not obstruct movement of upslope materials (Bjerrum 1967). Pore pressures and weathering processes acting over time are often required to reduce or eliminate properties controlling the shear strength of the clays. Carson (1977) concluded that the ease with which sensitive clays are disturbed anu remolded is important in determining the rate and extent of retrogressive failure and flow.

An example of a progressive failure in partly weathered clay given by Bjerrum (1967) is shown in Figure 49. Weathering of the overconsolidated clay has occurred to depth z, so that internal stresses (P_L) parallel to the surface are increased, water content has increased and shear strength has decreased to this depth (Fig. 49a). Toe erosion at the slope's base (Fig. 49b) has developed a steep slope and removed lateral support. Lateral stresses in the weathering zone are then transmitted to the lower unweathered clay by shear stress on the plane SS. These shear stresses combine with the shear stress from gravity to exceed the peak shear strength

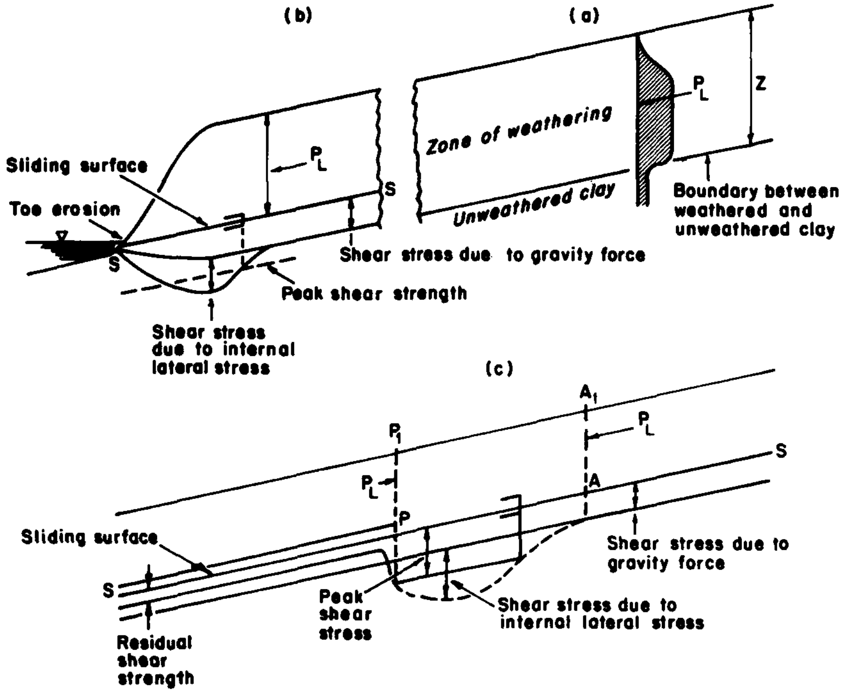


Figure 49. Example of a progressive slope failure in weathered clay. Mechanism is described in text (after Bjerrum 1967).

of the clay so that a slip failure on plane SS takes place and initiates the progressive failure.

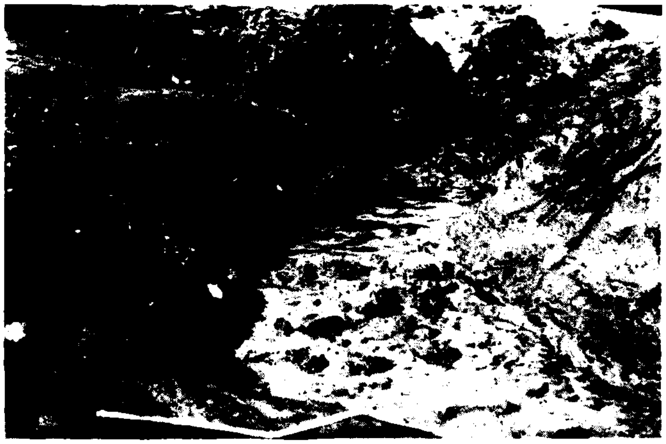
After the slip plane has formed (to point P in Fig. 49c), further movement depends on the slope inclination. Shear strength on the slip plane has been reduced to its residual value because it has moved. Further movement away from the remaining mass requires a slope angle sufficient to result in a gravitational force greater than the residual shear strength. This situation is shown in Figure 49c. Lateral stresses on plane PP_1 are therefore reduced. Once stresses on this plane are sufficiently reduced, shear stresses in the clay due to gravity at the leading edge of the slip surface (P_L) and the lateral stresses on the plane (AA_1) will exceed the peak shear strength, thereby initiating a second failure. Progressive development of the slip surface continues in the upslope direction.

Flow failures. The mobilization, movement and deposition of subaerial flows are complex and only partly understood. The sediment flow process has importance to shore erosion as a failure mechanism that is distinct from failure along a singular planar or circular surface, and as a downslope transport process which carries sediments ranging in size from clay to boulders into nearshore and offshore areas of the impoundment, possibly generating subaqueous flows or turbidity currents once within the water (e.g. Morgenstern 1967, Andresen and Bjerrum 1967, Hampton 1972). Flow failures of various sizes of river and coastal bluffs have been identified at a number of locations (e.g. Sharpe 1938, Varnes 1958, Chieruzzi and Baker 1958, Jones et al. 1961, Hutchinson 1970, Kachugin 1970, Bjerrum et al. 1971, McGreal and Craig 1977, Edil and Vallejo 1977, Heller 1981).

a. Viscous lobe of material on a slope of < 5° moving at a rate of about 1 m/hr (stadia rod is 2 m long).

b. View downslope of more fluid lobe of material moving at about 1 m/minute and with some sediment not fully assimilated into the flowing mass.

Figure 50. Examples of subaerial sediment flows with a matrix of clay- to sand-size material containing up to boulder-size clasts.



c. Debris flow moving in channel at a rate of about 0.5 m/s; coarse gravel clasts are being carried within the flow.

Figure 50 (con't). Examples of subaerial sediment flows with a matrix of clay- to sand-size material containing up to boulder-size clasts.

Commonly sediment flows are a part of complex or progressive failures with, for example, slip failures or liquefaction preceding and initiating remolding and the subsequent flow (e.g. numerous examples in Varnes 1978 and Skempton and Hutchinson 1969).

Although the process is simply referred to as sediment flow here, several different types have been identified. Each flow type has physical characteristics and apparent mechanisms of grain support and transport that distinguish them, yet they appear to actually represent a continuum of gradational forms. In mostly fine-grained materials, for example, they may exhibit behavior ranging from a very slow-moving, viscous, plastically deforming mass to a liquefied, almost fluid-like flow (Fig. 50) (e.g. Middleton and Hampton 1976, Youd 1973, Carter 1975, Lawson 1977, 1982b, Lowe 1979). Rates of movement can vary from centimeters per day to centimeters per second. As with lateral spreads, flows are commonly observed on slopes of 10° or less.

The mechanics of flow mobilization, the ability of certain flows to transport up to boulder size particles, and the mechanics of movement remain to be fully explained; recent theoretical and empirical treatments have significantly improved our understanding of the sediment flow process (e.g. Johnson 1970, Hampton 1972, Rodine and Johnson 1976, Keefer 1977, Takahashi 1981, Lawson 1982b). Observations and especially quantitative analyses of active subaerial flows and their properties have been strictly limited to date because of their general occurrence as singular, one-time events (e.g. Blackwelder 1928, Sharp and Nobles 1953, Curry 1966, Johnson and Rahn 1970, Johnson 1970, Rodine 1974, Pierson 1980), although Lawson (1977, 1982b) recently made detailed, repetitive measurements in the glacial environment where conditions are suitable for nearly continuous generation of subaerial flows during the summer.

The factors which apparently interact to produce conditions necessary for flow generation appear to be 1) rate and duration of precipitation, 2) geotechnical properties of slope material, including permeability and its variability with depth, 3) slope angle, 4) excess pore water pressures, 5) freeze-thaw activity, 6) slope aspect, 7) seepage pressures and ground water flow patterns, 8) snowmelt runoff, 9) vegetation cover, 10) thermal state and 11) stratigraphy of slope materials.

Of primary importance to the character of the sediment flow as well as to the initiation of movement is the initial process that directly causes loss of strength or remolding of the material, thereby reducing its shear strength and resistance to movement under the force of gravity. This process may involve, for example, reduction in cohesion or internal friction because of excess pore water pressures or leaching, the physical disaggregation and remolding of the sediments, or the disruption of particle contacts by earthquake motions. Water is inherently involved in loss of shear strength and flow mobilization and movement in most reported cases of subaerial flows in which fine-grained cohesive sediment is a major component (e.g. Blackwelder 1928, Sharp and Nobles 1953, Curry 1966, Crozier 1969, Rodine 1974, Keefer 1977, Lawson 1977, 1982b).

As movement takes place, various factors including changes in slope angle, turbulent mixing, addition of water and others may further reduce its strength. Deposition generally requires the opposite condition: increase in the strength or in the resistance to flow offered by the material (e.g. lowered slope angle). Within the shore zone, slow flows may undergo deposition at the base of the bluff, while more fluid and rapid flows may move directly into the impoundment.

A particular case of significance to bluff erosion is surface flow resulting from thawing of frozen bluffs in the spring (Edil and Vallejo 1977, Sterrett 1980, Reid 1983). Thawing causes melting of ice formed during freeze-up in the fall and winter. This meltwater can fully saturate or oversaturate the sediment, thereby reducing its shear strength. In addition, excess pore pressures may be generated under proper conditions above the still-frozen sediment, further reducing the strength of the materials (McRoberts and Morgenstern 1974a,b, McRoberts 1978). Thin flows of a few centimeters thickness characterize steeper slopes; thicker flows occur on lower angle slopes. Flow on frozen beaches has also been observed.

Subaqueous failures. Subaqueous slumping, from offshore and nearshore regions, including deltas, off coastal and lake shores has been described (e.g. Shepard 1955, Terzaghi 1956, Moore 1961, Morgenstern 1967, Coleman and Garrison 1977, Prior and Coleman 1978, Pickrill and Irwin 1983, Bea et al. 1983). Slope failures include localized minor slumps of fine-grained sediments mantling otherwise stable materials of relatively steep slope, intermittent slumping of recently deposited clays on gentle slopes, and movements encompassing a wide area of slope with flow and lateral spreading of fine-grained cohesionless material after failure by localized subsidence and translational motions (Terzaghi 1956). Liquefaction of bed materials by excess pore pressure from waves is also a possible failure mode (e.g. Gill and Nataraja 1983). Complex progressive failures similar to those described by Bjerrum (1967) or Skempton and Hutchinson (1969) were postulated to occur in shallow (5-to 25-m) water off the Mississippi Delta by Prior and Coleman (1978); the postulated movements and mechanisms are

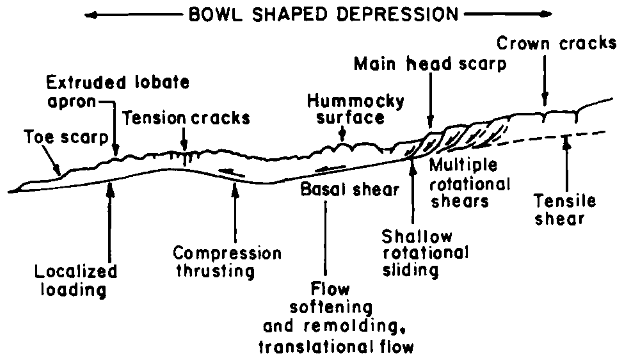


Figure 51. Postulated movements, mechanisms and configuration of subaqueous, progressive failures in shallow water off the Mississippi Delta (after Prior and Coleman 1978).

illustrated in Figure 51. Slopes in areas of subaqueous slumping are often low in angle (<10°) and have been reported as low as 1° to 3° (e.g. Shepard 1955). Sediments involved in failures on low angle slopes typically are normally consolidated or underconsolidated and fine-grained. Underconsolidated materials can originate by recent and rapid rates of deposition (Terzaghi 1956), producing material which is readily erodible and subject to failure by flow (Einsele et al. 1974).

Morgenstern (1967) analyzed the stability of subaqueous materials by using the limit equilibrium concept in terms of effective stress for drained and undrained failures, and considering the simple case of an infinite slope with slips along one or many closely spaced planes paralleling the slope surface. He also considers a third case termed collapse slumping for failure of metastable, underconsolidated sediments. This latter type results from failure initially under drained conditions but the deformations associated with failure cause a large and sudden increase in pore pressures. This increase in turn reduces shearing resistance and accelerates the moving sediment mass.

The stability analyses follow those for drained and undrained failures of saturated materials discussed previously. Drained failures are probably limited to coarse-grained (sand, gravel) materials on steep slopes. Undrained failures are probably typical of underconsolidated materials or those where stresses are induced by rapid deposition or erosion.

Additionally, Morgenstern (1967) considers the effects of underconsolidation on undrained strength, which he deduced should be proportional to the average degree of consolidation. Thus,

$$\left(\frac{c_u}{p_m}\right)_{\bar{u}} = N\bar{u} \tag{34}$$

where $\bar{u}$ is the average degree of consolidation, p_m maximum effective overburden pressure, and

$$N = \frac{c_u}{p_m} \quad . \tag{35}$$

This relationship rests on the assumption that effective overburden pressure p at any time during consolidation when excess pore pressures exist is given by

$$p = \gamma' z - u = \gamma' z \left(1 - \frac{u}{\gamma' z}\right) \quad (36)$$

where u is excess pore pressure, γ' is submerged unit weight of soil, and z is depth. Excess pore pressure can be estimated as varying linearly with depth:

$$u = nz \ . \quad (37)$$

Substituting in eq 36 gives

$$p = \gamma' z \left(1 - \frac{n}{\gamma'}\right) \ . \quad (38)$$

This relationship shows that excess pore pressures can develop in material undergoing an increase in height due to deposition (Terzaghi 1956). The pore pressure values depend upon rate of sedimentation, height of the deposit, and coefficient of consolidation for the material. At any depth in the material, u will reduce the effective stress and undrained strength of the material. Clearly, failure conditions can develop at some depth over time when consolidation does not keep pace with rates of sedimentation.

Factors that might lead to subaqueous slope failures in reservoirs remain speculative, but several situations may be conducive to failure. First, oversteepening of nearshore sediments in slopes can result from erosion by wind waves and currents. This may be particularly true when the combined effects of erosion and wave-generated pore pressures in submerged sediments are at a maximum during storms (e.g. Henkel 1970, Suhayada et al. 1976, Tsui and Helfrich 1983).

Similarly the failure of bluff slopes and the deposition of this sediment mass upon subaqueous slopes could increase the overburden sufficiently to produce an unstable condition in slopes of low angle. This situation is analogous to that described by Hutchinson and Bhandari (1971) for undrained loading of subaerial slopes (Fig. 52). Rapidly eroding and receding shores can introduce a large quantity of sediment into the reservoir pool. If it is deposited rapidly on nearshore slopes, a metastable condition may exist

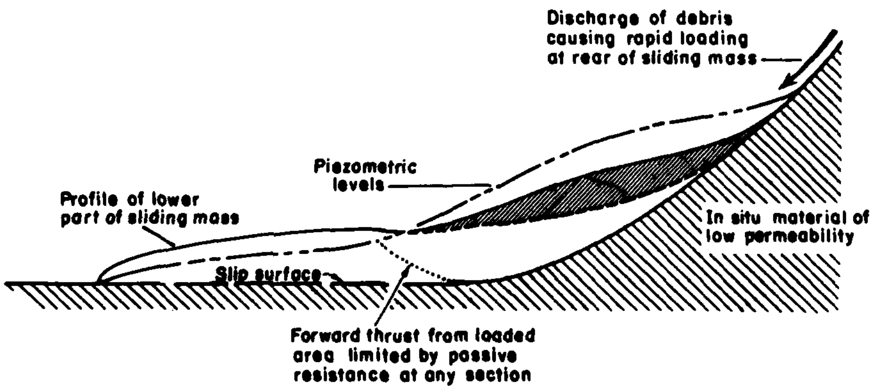


Figure 52. Schematic illustrating common development and movement of slopes due to undrained loading (after Hutchinson and Bhandari 1971).

and collapse slumping as described by Morgenstern may result from continuing sedimentation. Slides in shallow water may result from rapid downdraw or the direct action of waves and nearshore currents, particularly shallow types affecting the transitory part of the shore zone. Lowering of water level may similarly result in movement of deposition by incoming rivers onto delta slopes and cause oversteepening and deep- or shallow-seated failures of those slopes (e.g. Pickrill and Irwin 1983). Because of unfavorable environmental conditions, observations and measurement of subaqueous failures remain to be done.

Subaqueous failures can also generate or result in subaqueous retrogressive flow slides, debris flows, turbidity currents, liquefaction followed by debris flow, grain flows and others (e.g. Terzaghi 1956, Morgenstern 1967, Andresen and Bjerrum 1967, Hampton 1972, Carter 1975, Lowe 1976, Middleton and Hampton 1976). Initiation of movement likewise results from a loss of strength or resistance to shearing under the force of gravity, such as may result from temporary increases in pore pressures, shock (from earthquakes or perhaps sudden mass loading), effective oversteepening of sediments in slope, or perhaps fluidization resulting from upward flow of ground water through the bottom sediments (Carter 1975).

Stability factors

Factors identified as critical to the stability of subaerial and subaqueous slopes include: 1) ground water conditions, 2) stratigraphy with respect to bluff orientation, 3) occurrence of potential "weak" layers or failure surfaces, 4) intensity and type of toe zone erosional processes, 5) intensity and type of bluff face erosional or degradational processes, 6) slope geometry (mainly height, length, angle, and aspect), 7) geotechnical properties of the sediments and their variabilty within composite slopes, 8) nearshore bottom topography, and 9) climate/weather.

One factor deserving further discussion that is clearly important and interacts with the other factors to cause shallow or deep-seated instability is the presence of water. Field and theoretical analyses have indicated that water is usually critical in determining or modifying the shear strength of slope materials and thus their frictional and cohesive resistance to the force of gravity. As saturation increases, the simple increase in the mass of slope materials effectively increases the applied shear stress (e.g. Terzaghi 1950). The horizontal movement of water generates seepage pressures that generally reduce stability; concentrated flow within single layers or along fracture planes will locally reduce the effective stress and lead to slippage (e.g. Rodgers and Selby 1980, Sterrett 1980). High flow conditions can result in springs issuing at a bluff face that may in turn cause piping and undermining of overlying sediments (e.g. Hadley 1976, Hagerty et al. 1981, Hopkins et al. 1975).

Seepage into submerged sediments may actually increase the stability of these materials in accordance with the average hydraulic gradient (e.g. Burgi and Karaki 1971, Thomson and Morgenstern 1977), and decrease their erodibility, in part because of the deposition of a silt seal within particle pores from suspended fines as water enters them (Harrison 1968, Harrison and Clayton 1970). Conversely, outflow may increase their erodibility by decreasing the effective cohesion and hence shear strength (Terzaghi and

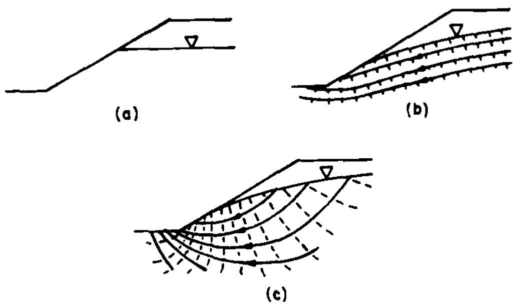


Figure 53. Ground water flow systems in slopes: (a) static case typically assumed in slope stability analyses, and (b) commonly assumed, but incorrect, parallel to slope flow pattern. Actual flow system in homogeneous isotropic materials (c) should be used in predicting pore pressure distributions along potential slip surfaces. Flow systems in nonhomogeneous and anisotropic materials are more complex than shown (after Patton and Hendron 1974).

Peck 1967), possibly leading to heaving and failure (Terzaghi 1929). Outflow may reduce bed material strength and increase erodibility by currents (e.g. Clayton et al. 1966).

Weather-related factors may be equally important in affecting stability and sediment strength. These include rate and duration of rainfall, rate and volume of snowmelt, and extended periods without rainfall. Drying periods followed by wetting during heavy precipitation may result directly in sloughing (Quigley et al. 1978, Kachugin 1970). Long-term ground water flow may also leach chemical constituents from sediments and reduce strength (e.g. Kachugin 1970).

The ground water flow regime is thus critical to analysis of shore zone stability and should be defined by field measurement, as the actual flow pattern may deviate significantly from a typically assumed parallel-to-slope flow regime or from an assumed hydrostatic condition (Patton and Hendron 1974, Cedergren 1977, Hodge and Freeze 1977, Lafleur and Lefebvre 1980) (Fig. 53). Regional geology (stratigraphy and associated regional ground water flow pattern) may be critical to defining the actual, more localized pore pressure distribution in a slope, as well as potential problems in slope stability that are not evident at the ground surface. That is, slope materials may undergo progressive changes in strength by leaching, piping or other ground water processes and cause an unexpected failure.

Factors affecting shear stresses and shearing resistance

Perhaps because of the relative simplicity of slip failures, they have generally received more attention than other types of slope movements. Hence, many of the factors determining the stability of slopes with regard to slip failure have been identified. Factors leading specifically to slip movements in bluff slopes are listed below, based upon a compilation by Varnes (1978) of 1) factors that contribute to an increase in shear stress and 2) factors that reduce shear strength:

Factors increasing shear stress

1. Removal of lateral support by wave and current erosion, previous slope failures, and surficial degradation by weathering, wetting and drying or frost action.

2. Surcharge on slopes due to weight of precipitation, failed material from an upslope position, seepage pressures or vegetation.

3. Earthquake shocks and other human-induced vibrations.

4. Removal of underlying support by wave and current undercutting, subaerial weathering, wetting and drying, frost action, piping, or failure of underlying materials.

5. Lateral pressures as may result from water in cracks, ice formation in cracks or soil pores, or mobilization of residual stress.

Factors reducing shear strength

1. Inherent characteristics of material, such as composition, internal structure, geologic discontinuities such as bedding planes, joints or fractures, massive materials on weak beds, alternating permeable and impermeable strata, and slope orientation.

2. Weathering and related physical and chemical reactions that may include physical disintegration by frost or thermal expansion, hydration of clay minerals, cation exchange in clay minerals, drying and cracking of clays, or solution of cementing agents.

3. Changes in intergranular forces due to changes in water content and excess pore pressures within sediment or fractures and other discontinuities.

4. Structural changes caused by fissuring of clays, spalling with removal of surficial materials, disturbance or remolding of fine-grained materials, burrowing animals, and growing tree roots.

OVERLAND FLOW

The subaerial erosion of unconsolidated slopes by overland flow processes is complex, involving relationships and mechanisms such as those of erosion by streams (e.g. Fenneman 1908, Bryan 1922, Lawson 1932). Although the importance of overland flow processes in eroding and shaping slopes was identified early on, their significance as a modifier of inland shores and streambanks has only more recently been expounded upon (e.g. Chieruzzi and Baker 1958, Brunsden and Kesel 1963, Pezzetta and Moore 1978). An understanding of the overland flow processes is necessary to understand how and

why certain bluffs recede while others adjacent to it do not and the overall importance of these processes to shore zone erosion.

In general, net erosion by overland flow processes depends upon the magnitude of the erosional process, the ease with which soil particles are detached, the capacity of the process for transporting erosional products from the slope, and the ease of entrainment of the soil particles (Ellison 1947a,b). Four general sub-processes of erosion cause soil loss on slopes: 1) detachment by rainfall, 2) detachment by runoff, 3) transport by rainfall and 4) transport by runoff (e.g. Foster and Meyer 1972). Both laminar and turbulent runoff flows may take place, with hydraulic and geologic parameters defining which occurs (Emmett 1970, Savat 1977, Moss and Walker 1978). Many physical, chemical and compositional properties combine to determine the erodibility of sediments by rainfall and runoff processes.

Overland flow processes will be discussed in this section in terms of raindrop splash, sheet flow, rill flow, and gully development and flow. Some authors also consider creep, which is an especially important modifier of lower angle slopes, in the same class as these processes (e.g. Schumm 1956, Culling 1963, Kirkby 1967). Creep is, however, considered in this monograph as a variety of mass movement, being a slow, more or less continuous downslope movement of sediment under the force of gravity.

As with many complex surficial processes, certain fundamental aspects of overland flow processes are not well understood. For example, Shen's (1979) review of the literature indicated that primary factors (such as soil moisture) which affect soil erodibility and the basic mechanisms of overland flow transport are known only qualitatively. Recently, Zaslavsky and Sinai (1981) have questioned several fundamental concepts of surface hydrology, especially the role of seepage in causing erosion by rainfall and causing runoff on initially unsaturated soils. Literature on overland flow processes is scattered widely through agricultural, engineering and geological sources and can only partly be treated in this report; the reader may consult Zachar (1982, p. 205-347), Vanoni (1975, Chap. 4) and other references cited here for further discussion of erosion and other aspects of sedimentation resulting from overland flow processes.

Raindrop splash

Raindrop impact and transport are only partly understood because of an inherent natural variability in the intensity, dimensions and angle of incidence of raindrops during and between individual storms.

The primary effect of rainfall is in causing the disaggregation and detachment of individual particles and less often small cohesive aggregates, and thus in loosening and dispersing the uppermost mineral soil horizon or sediments. Detachment results from the kinetic energy released upon raindrop impact (Laws 1941, Laws and Parsons 1943, Ellison 1944, 1947a,b, Ekern 1950, Wischmeier and Smith 1958, Mutchler and Young 1975). The net result on the materials is to reduce shear strengths by effectively decreasing particle packing and effective cohesion, thus making them more susceptible to entrainment by sheet flow or other slope processes.

Individual particles mostly move downslope some distance following raindrop impact, although a certain smaller percentage will actually move

upslope (Ellison 1944). The amount of displacement is termed raindrop erosion or rainsplash movement and apparently ranges from a few millimeters to a few meters (Zachar 1982). The combined effects of disaggregation and movement are independent of sheet flow or other overland flow processes (Cook 1936). Overall, however, raindrop impact without runoff is not believed to be effective in causing erosion (Emmett 1970).

An estimate of the amount of sediment moved by raindrop impact was defined by Ellison (1947a,b) based upon the kinetic energy of falling rain:

$$E = k\, v^{4.33} \cdot d^{1.07} \cdot i^{0.65} \qquad (39)$$

where E = the amount of soil splashed per interval of 30 minutes
v = drop velocity
d = drop diameter
i = rainfall intensity (in./hr)
k = a constant determined by soil type.

Empirical equations exist for calculating drop velocity and average rainfall intensity. Tables have also been prepared for drop size diameters based upon rainfall intensity (see Zachar 1982).

Mirtskhulava (1970, in Zachar 1982) empirically defined the actual amount of eroded material by rainsplash (q_D), based upon field measurements, as

$$q_D = t \left(\frac{0.13\ \gamma\ i\ v_c^{\,2}}{2g}\right) \left(\frac{d_K v_c}{d_{dk}\ v_{cd}}\right) - 4 \sin \alpha \qquad (40)$$

where γ = density of saturated soil
i = rainfall intensity
v_c = terminal velocity of raindrops
t = duration of rainfall
d_k = mean drop diameter
d_{dk} = limiting size drop below which erosion does not occur
α = angle between vertically falling rain and the ground surface
g = acceleration due to gravity.

This equation assumes a limiting raindrop size below which no erosion occurs. Kneale (1982), however, recently conducted field analyses of this relationship and concluded that, for the sandy loam soil tested, measurable amounts of rainsplash movement took place at even the lowest rainfall intensities measured (less than 5 mm/hour). Kneale suggested that such a lower limit may not exist under natural conditions because drop-size distribution has little influence on the overall energy and momentum of the total rainfall mass at low intensities.

The amount of surface erosion resulting directly from rainsplash on a sloping surface can be significant (Zachar 1982), particularly in regions with high intensity rainfall. But rainsplash is more important in modifying the structure and strength of the materials, thus increasing their susceptibility to erosion by flowing water (Smith and Wischmeier 1957). At one location, for example, the weight of particles of bare soil moved by

rainsplash during a cloudburst was estimated at over 97 tons per acre for an average precipitation rate of 0.8 in./hr (Osborn 1954a). According to Meyer (1971), in most parts of the United States each square mile of land is bombarded by several quadrillion raindrops annually. For a region with 30 in. of rainfall annually, the impact energy is approximately 30 billion foot-pounds or the equivalent of 10,000 tons of TNT. Thus, tremendous quantities of soil particles can be detached and moved about.

Bare areas are therefore especially susceptible to splash erosion, even when such areas occur within an otherwise vegetated area, including forests, and may suffer 1 to 2 orders of magnitude greater erosion (e.g. Osborn 1954a,b, Schumm 1956, Karaushev et al. 1974, Imeson 1977). The density and cover of vegetation are thus extremely important factors. Additionally, the effectiveness of raindrop erosion depends upon the geotechnical properties of the exposed sediments and upon the angle and aspect of the slope with respect to prevailing winds and storm tracks. Hail also augments and significantly increases the impact effects of precipitation (Zachar 1982).

In addition to surface erosion and runoff, a subsurface flow may take place that can transport fine-grained products of erosion. Pilgrim and Huff (1983) concluded that rainwater flowing through macropores in soil may transport fine-grained particles in suspension. They found that sediment of very fine silt size (4- to 8-μm diameter) on a field plot was carried in suspension by subsurface runoff through small cracks, root holes, animal holes or other soil macropores. Concentrations were similar to those of moderate stream flows and commenced shortly after rainfall began. Pilgrim and Huff concluded that most, if not all, of the suspended material was actually detached and entrained at the ground surface by raindrop impact. Thus, the removal of some sediment from slopes may take place even without surface runoff by sheet or rill flow.

Sheet flow

Often enough rain falls to produce a surface flow of water in thin sheets which will transport detached particles down a slope. The amount of rain necessary for this to occur is as yet very difficult to define and varies across the same slope or from site to site.

Similarly the transition from sheet flow to concentration within rills remains problematic. Why rills develop at some locations but not others cannot yet be defined (e.g. Emmett 1970).

In addition to rainfall rates, a factor important to both sheet and rill flow is the infiltration capacity of the slope materials (e.g. Schumm and Lusby 1963). Infiltration capacity is the maximum sustained rate at which a soil (sediment) will transmit water; hence, if exceeded by rainfall, water begins collecting on the surface and shortly thereafter coalescing in sheet flow. Soils with a low infiltration or storage capacity exhibit a rapid initiation of sheet flow. Surface roughness, slope dimensions and slope geometry will likewise affect flow initiation. Frozen, ice-rich soil represents a common, but extreme, case of exceedingly low or nonexistent infiltration capacity (e.g. Dingman 1975, Zuzel et al. 1982).

Sheet flow has two major effects. One is the physical washing or removal of loose particles or small aggregates at the ground surface. The dimensions of such particles are determined by the ability of the water to move them, with only fine-grained sediments moving during low intensity rainfalls. Entrainment of material by sheet flow follows the hydraulics of streamflow and thus a certain amount of tractive force is required, although rainsplash and other processes may loosen or detach particles and facilitate transport. Young and Mutchler (1969a,b) and Emmett (1970) found that rates of removal decreased with time because these more easily eroded particles are removed quickly, and flows then possess insufficient force to erode embedded particles. As Foster and Meyer (1972) pointed out, the rate of sediment transport is the lesser of either the flow's transport capacity or of the rate at which particles become available for transport.

The second major effect of sheet flow is chemical and mainly involves the leaching of easily soluble substances or the removal of organic matter from the uppermost sediment or soil. Fertilizers and biocides are the source of the soluble substances and are important as potential contaminants in the impoundment and in downstream waters.

Particles or aggregates transported by sheet flows have usually been detached by raindrop impact before a thin water layer develops on the surface or by other processes, such as frost action, wetting and drying, impact of hail, churning by animals and insects, erosion by flowing water, and longer term weathering processes. Cryogenic processes are particularly important in winter and spring. According to Zachar (1982), the movement of particles is greatest when frozen soil is thawing during a heavy rain, the quantity increasing in proportion to the kinetic energy of the rain. As slope angle and length increase, the potential for mechanical erosion by flowing water (rather than a simple washing of previously loosened particles) increases (Zingg 1940), but this appears to be much less important than washing and only occurs as rill formation begins (Moss and Walker 1978).

Downslope concentrations of sheet flow can result from obstacles such as clumps of vegetation or rocks in the flow path that increase the water depth and velocity of flow downslope of their location. Raindrops are also important because their impact on the water surface can create turbulence and increase resistance to flow, in turn decreasing flow velocity and increasing flow depth (Emmett 1970). The combined effects of rainfall and runoff apparently increase transport of fine-grained sediment but decrease it for coarse-grained sediment (Meyer and Monke 1965).

Additional factors affecting sheet erosion include the angle of the slope, length of slope, surface roughness, vegetative cover and soil (sediment) type and properties (e.g. Smith and Wischmeier 1957, Schumm 1956, Meyer and Monke 1965, Lyle and Smerdon 1965, Young and Mutchler 1969a,b, Mutchler and Greer 1980). The effects of these factors will be discussed in more detail in the section Overland Flow Factors.

Snow thaw erosion

As Zachar (1982) points out, the erosion caused by snowmelt runoff, which he refers to as snow thaw erosion, can be considerable on slopes as

gentle as 2° or so, especially in regions characterized by heavy snow precipitation and sudden thawing. It is, however, not always accounted for in considering sediment or soil losses from slopes.

In this same sense, slopes with frozen soil and sediment in association with a minor snowmelt or precipitation event may locally produce major erosional losses (e.g. Tigerman and Rosa 1949, Atkinson and Bay 1940, Zuzel et al. 1982). Zuzel et al. (1982) examined erosion in an area of intermittent snow accumulation where several cycles of snowfall and snowmelt runoff took place during winter, each causing a significant loss of soil that exceeded erosion for the remainder of the year. This significant erosion is controlled by the effects of freeze-up during fall, the amount of snow that melts and thus the quantity of runoff, and the presence of frozen ground beneath a surficially thawed layer. As with rainfall effects, the length and angle of slope, moisture content of the soil, and the cover of vegetation also affect erosion.

During freeze-up, alternating periods of freezing and thawing can cause heaving of the surficial materials and thus loosen and detach soil aggregates or particles (Schumm and Lusby 1963). In addition, final freeze-up results in an upward migration of water within the sediments toward the downward advancing freezing front. As a consequence, ice crystals grow which exceed available pore space, heaving and loosening the upper frozen materials. On bare slopes exposed to direct sun in winter, sublimation of the ice crystals releases soil particles and aggregates which may fall or roll to the slope's base if sufficiently steep (e.g. Harrison 1970).

Upon thawing, two effects result: one is the freeing of this loosened material to be eroded by flowing water or raindrops, and the second is to release excess quantities of pore water which oversaturate the thawed sediment, making it susceptible to failure and downslope flow (e.g. Atkinson and Bay 1940, Bay et al. 1952, Tigerman and Rosa 1949, Schumm and Lusby 1963, Soons and Rainer 1968). The presence of frozen ground beneath this thawed layer greatly reduces the infiltration capacity of the sediment and effectively accelerates runoff, whether from precipitation or snowmelt (Zuzel et al. 1982, Zachar 1982). Bare slopes of southern aspect are particularly susceptible to the combined effects of freeze-thaw cycles, snowmelt and surficial thaw.

Rill and gully erosion

The cause of rilling (Fig. 54) is apparently a function of the infiltration capacity of the sediments in relation to the quantity of runoff generated by precipitation or snowmelt, the slope length, and the slope steepness. At the outset of rainfall, some water is consumed in wetting the surface and filling pore spaces in the upper soil while disaggregation of the surface crust and splashing by raindrops takes place. Subsurface runoff within soil pores may be occurring during this time (Pilgrim and Huff 1983). Eventually the uppermost sediments become sufficiently saturated to decrease the infiltration rate and increase surface runoff.

The depth of flow in sheets and rills varies downslope. This depth varies downslope in direct proportion to the length of the slope x, intens-

Figure 54. Rills developed on a relatively steep slope composed of glacial deposits, principally till. Slope height shown is about 10 m. Note increase in rill development downslope.

ity of runoff q, and Manning's surface roughness n, and in inverse proportion to the angle of slope α (Horton 1945):

$$d_x = \left(\frac{q_r \; nx}{1020}\right)^{3/5} \frac{1}{\tan^{0.3}\alpha} \quad . \qquad (41)$$

Rill development is significantly affected by the material composing the slopes, particularly its permeability and storage capacity, and its resistance to erosion by flowing water. Composite slopes exhibit complex rill development. On slopes composed of permeable material, rill development is discontinuous and highly irregular in form (Barendregt and Ongley 1979); erosion on these slopes appears mostly to result from rainsplash and sheet flow (Zachar 1982). On less permeable materials, rill erosion is significantly more important. Rilling then depends upon the resistance of the materials in relation to the tractive force exerted by the flowing water. For example, the resistance to erosion varies linearly with its degree of compaction as well as properties affecting the shearing resistance of the sediment (Lyle and Smerdon 1965).

Figure 55. Gully in slope composed of glacial deposits and located below small surface depression characterized by runoff flow during spring snowmelt.

Entire slopes may be crossed by a dense pattern of rills, which tend to flow straight downslope and may eventually join into wider and deeper ones (Fig. 54). Rills are thus distinct from gullies, which are similar to small valleys or ravines with steep sides and often curvature of the channel bed but which tend to occur sporadically across the surface. Water in both rills and gullies is present only during periods of precipitation or snowmelt runoff.

Gullying of slopes typically results from concentrated runoff generated on adjacent upland areas, such as tilled agricultural fields (McComas 1974), or from the downslope coalescence of rills (Fig. 55). Typically gullies are found at a regular spacing on bluff faces when developed from rills (e.g. Brunsden and Kesel 1973, Zeman 1978, Sterrett 1980). Gully deepening and widening may result both from mechanical erosion of the channel bed by flowing water and from sidewall processes such as fluvial undercutting and failure, surface flow from intergully highs, rainsplash, frost action, desiccation, and wind erosion (e.g. Piest et al. 1975, Zeman 1978, Blong et al. 1982). Products of these processes are removed from the gully floors by flowing water.

Rates of gully erosion and enlargement often far exceed the rate of bluff retreat as a whole, resulting in scalloped areas on bluff faces. Zeman (1978) estimated rates up to 10 times as rapid in the gullies as in the actual recession of blufflines. Extremely rapid rates also result from overflow of water within the gully and cause retrograde erosion and enlargement of the gully width (Zachar 1982). The resistance of the sedi-

ments to erosion by flowing water and to sidewall processes is clearly an important factor governing gully formation.

Flow mechanics

The rate of erosion depends principally upon the velocity of the runoff flow and upon the erodibility or resistance of the soil relative to tractive forces exerted by the flowing water. Flow velocity (v) depends on the runoff rate (q), the slope steepness (s) and the hydraulic roughness (n) of the flow. Meyer (1965) and Meyer and Wischmeier (1969) concluded that the velocity of flow in rills on agricultural lands is proportionally related to these factors by

$$v \propto \frac{s^{1/3} q^{1/3}}{n^{2/3}} \tag{42}$$

As an estimate, Meyer (1971) pointed out that the tractive force exerted by flowing water, and the transport capacity of flowing water, are proportional to v^2 and v^5 respectively, indicating that even small changes in s, q or n can induce significant changes in the erosion rate within rills or gullies. The factor n usually ranges from 0.1 to 0.05.

In a similar relationship, the velocity of flow of water in rills and gullies has been considered analogous to flow in ephemeral streams. Flow velocity can therefore be estimated by using the standard, empirically based, Manning equation for river flow (e.g. Leopold et al. 1964, Young 1972):

$$v = 1.49 \frac{R^{2/3} s^{1/2}}{n} \tag{43}$$

where s is slope angle, n is hydraulic roughness and R, the hydraulic radius, is defined by

$$R = \frac{A}{P_w} \tag{44}$$

with A the area of the channel cross section in square feet and P_w the length of the wetted perimeter in feet. Emmett's (1970) data suggested a typical n-value of 0.05.

Savat (1977), however, questioned this traditional usage for sheet flow because of the flow depths involved and, based upon experimental data, concluded that this formula best applied only to transitory conditions between laminar and turbulent flow. Savat's data suggested that purely laminar and purely turbulent sheet flow (which is common on steep slopes) are represented, respectively, by the Horton et al. (1934) formula of

$$v = 327\ d^2 \sin s/\nu \tag{45}$$

and by the Darcy-Weisbach formula

$$v = \frac{8g\ R \sin S}{0.053} \tag{46}$$

where d is depth of flow above bed and ν is kinematic viscosity. Coefficient values are based on Savat's experimental results.

Water flowing in both rills and gullies erodes their beds and transports sediment as both suspended load and bed load. Recent observations by Moss and Walker (1978) indicated that flow is almost always turbulent and supercritical within rills, with flows nearly always loaded to capacity with sediment. Even very shallow sheet flows that are not supercritical or turbulent readily entrain and transport sediment without the aid of raindrop impact (Emmett 1970, Foster and Meyer 1972).

Water as shallow as a millimeter transports particulate matter in suspension and at the bed by rolling and saltation or intermittently in traction (Meyer and Monke 1965, Emmett 1970, Moss and Walker 1978). Most fine-grained suspended sediment of silt and clay size moves directly into adjacent streams (Ellison 1947a,b); these sizes are critical to transport of organic and inorganic nutrients. Coarser particles are deposited on lower angle slopes or remain in the channel bed as flow subsides. Particles up to pebble size were observed by Moss and Walker (1978) to be moved as bedload within rills on slopes as low as 4° with sufficient (but not exceptional) discharge. Suspended load transport was controlled by the availability of fine-grained particles and thus, if available, full capacity of the flow was readily attained.

Carson (1971) concluded that data from White (1940) indicated that suspension of sediment is initiated by a threshold shear velocity $u*_c$, defined by

$$u*_c = \frac{1}{7}\,\omega \tag{47}$$

where ω is the terminal fall velocity of the particles.

On lower angle slopes of 1° to 4°, bedload transport is also active, but limited usually to sand-size particles that undergo saltation, to fine gravels in traction, and to particles 10 mm in diameter or less that roll along the bed, some even protruding above the water surface (Leopold and Miller 1956, Moss and Walker 1978).

Plant debris, such as roots or twigs, often is floated or dragged along within rills and gullies. If such material becomes lodged across them, a small dam forms, which locally develops a pool behind it and causes scour below it.

Active scour is apparently limited to rills and gullies without channel deposits, so that the underlying slope sediments are exposed directly to fluvial action. Moss and Walker (1978) observed that in gullies and rills in which a gravel lag (armoring) remained on their beds, the gravel prevented scour but also apparently shifted erosion laterally to channel walls. This eventually led to widening the gully to form of a deep, steep-sided ravine. It also prevented further removal of fine material as suspended load.

As with rainfall, vegetation cover strongly inhibits erosion by surface runoff, but any small bare areas within an incomplete vegetation cover are still subject to rapid erosion. Plants cause an increase in resistance to flow which may induce deposition of sediment load, dissipate the energy of raindrops, and reduce the boundary sharpness between the fluid and bed, thus making surface flows deeper, slower and less turbulent than on bare ground.

The erosion of sediment by flowing water depends upon the tractive force or drag exerted on particles in the channel bed, and upon a net upward lift force that acts from the bed and near the fluid boundary. This lift force, while often neglected in estimating the critical shear velocity at which sediment motion begins, may actually equal the tractive force at the bed but falls off rapidly just above it (e.g. Watters and Rao 1971). Lift force is capable of initiating transport irrespective of the tractive force magnitude (Carson 1971, Allen 1982). Both forces increase with the square of the velocity and can be expressed for steady channel flow in the following form:

$$F_D = C_D a \frac{\rho_f}{2} u_o^2 \tag{48}$$

and

$$F_L = C_L a \frac{\rho_f}{2} u_o^2 \tag{49}$$

where F_D = the drag force
F_L = lift force
a = the surface area of the particle
u_o = the velocity of fluid around the particle
C_D = a proportionality constant, a function of particle shape, termed the drag coefficient
C_L = the lift coefficient (Carson 1971).

Alternatively, Allen (1982) derived an equation for $\tau_{o(cr)}$ for both drag and lift forces in streams, assuming $F_L = K F_D$, where k is a coefficient varying consistently with flow conditions determining F_D:

$$\tau_{o(cr)} = \frac{2}{3} (\gamma_s - \gamma)\ gd \left(\frac{\sin \alpha}{\cos \alpha + K \sin \alpha}\right) \tag{50}$$

where α = angle of interlock between particles. In nondimensional terms and with α expressed using d, d_o and s, eq 50 becomes

$$\frac{\tau_{o(cr)}}{(\gamma_s - \gamma)\ gd} = \frac{2}{3} \left\{ \frac{\frac{1}{2}(d_o + s)}{\left[\frac{d^2}{4} + \frac{dd_o}{2} + \frac{d_o s}{2} + \frac{s^2}{4}\right]^{1/2} + \frac{k}{2}(d_o + s)} \right\} \tag{51}$$

where d_o is the average particle diameter on the bed and s the distance separating bed particles. Thus threshold stress increases with particle diameter d, ratio of d and d_o, separation distance s, and the density difference. The relative importance of the two erosional forces in shallow overland flows requires additional research.

AD-A157 811 EROSION OF NORTHERN RESERVOIR SHORES: AN ANALYSIS AND 2/3
APPLICATION OF PERTINENT LITERATURE(U) COLD REGIONS
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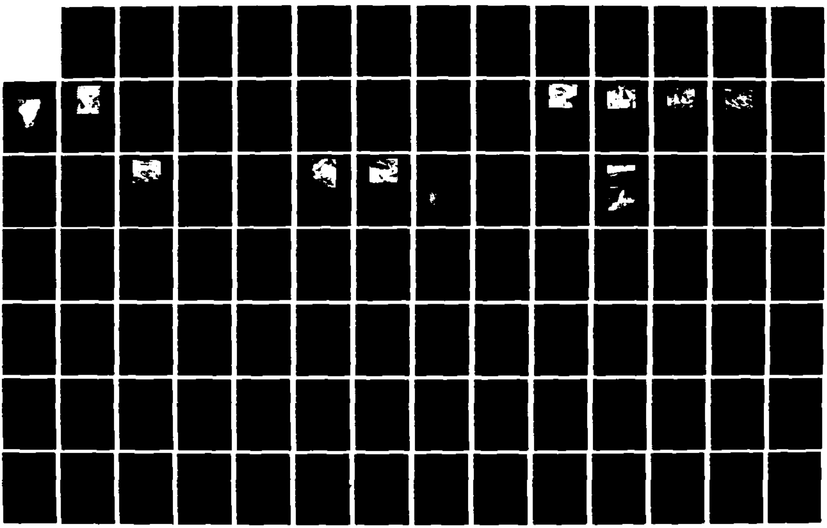
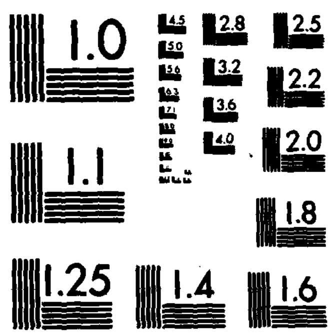




MICROCOPY RESOLUTION TEST CHART
DARDS-1963-A

In general usage, estimates of discrete particle motion at a stream bed have typically been determined by computing an average tractive force τ_o and assuming uniform flow in a straight channel. Thus,

$$\bar{\tau}_o = \rho g R s \qquad (52)$$

where ρ = fluid density
g = acceleration due to gravity
R = hydraulic radius
s = water surface or channel bed slope (e.g. Carson 1971, Simons and Li 1982).

For a more complex situation with a nonuniform water surface but mostly uniform bed profile, tractive force is better approximated by

$$\bar{\tau}_o = \rho g\ Rs + \rho \frac{\bar{u}^2}{P} \frac{dA}{dx} \qquad (53)$$

where A = channel cross sectional area
x = distance along the channel
$\bar{u}^2$ = mean square velocity
P = the wetted perimeter (Bathurst 1982).

If the bed surface is also nonuniform, as in a series of riffle and pool sequences in a gully crossing a slope composed of sediments of differing erodibility, neither equation applies. At high discharges, however, both equations begin to more closely approximate bed shear stress, presumably because the pool/riffle sequence becomes less influential as flow depth increases. In general, the flow depth, cross-sectional profile and thus available force to cause erosion are a function of 1) rainfall intensity, 2) infiltration capacity, 3) length of overland flow, 4) slope angle, 5) surface roughness and 6) degree of turbulence.

Critical shear stresses exerted on the walls or banks of rills and gullies can also be estimated by first determining the critical shear stress at the bed and then reducing it by a factor K defined by

$$K = \frac{(\bar{\tau}_s)_c}{(\bar{\tau}_b)_c} = \cos\theta \left(1 - \frac{\tan^2\theta}{\tan^2\phi}\right)^{1/2} \qquad (54)$$

in order to account for the force exerted by the downslope component of weight acting on any particle in the sloping channel sides (Simons and Li 1982). In this equation, θ is the angle of side slope and ϕ is the angle of repose of bank sediments. A chart for use in calculating $\bar{\tau}_s$ and $\bar{\tau}_b$ used by Simons and Li (1982) for stream channels of differing shapes is shown in Figure 56. For example, according to this relationship, the maximum value for the bank shear stress in a wide channel (i.e. width much greater than depth) is

$$\bar{\tau}_s = 0.77\ \bar{\tau}_b \qquad (55)$$

The critical tractive force or critical erosion velocity required to initiate bed movement varies with the grain size and certain properties of

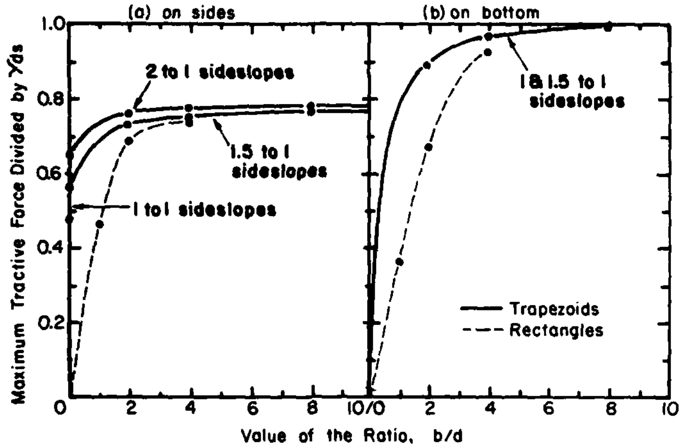


Figure 56. Chart for estimating maximum tractive force on sides of channels of trapezoidal or rectangular cross section (after Simons and Li 1982).

the bed material. Numerous investigations of this relationship have been made (e.g. Hjulström 1935, Shields 1936, White 1940, Bagnold 1954, Sŭndborg 1956, Yalin 1977, see Allen 1982 for a comprehensive reference list).

Typically the threshold of motion for cohesionless material by some critical tractive force is estimated with the criteria defined by Shields (1936) for a critical boundary shear stress. Shields used an empirical relationship between a dimensionless boundary shear stress, called the entrainment function,

$$F_* = \frac{(\tau_o)_{cr}}{(\gamma_s - \gamma)\, d} \qquad (56)$$

and the conditions of flow using the grain boundary Reynolds number,

$$R_e = \frac{u_{*(cr)}\, d}{\nu} \quad . \qquad (57)$$

In eq 57 $u_*(cr)$ (the critical shear velocity at the moment of particle motion) is equal to $\sqrt{\tau_o/\rho}$ (where ρ is fluid density, ν is kinematic viscosity and d is particle diameter). R_e characterizes the dynamic stability of flow near the bed in relation to particle stability. Values can be estimated from a graph of the Shields' criterion (Fig. 57). It should be noted that this relationship applies to the initial movement of grains on a plane bed composed of well-sorted or uniform grain sizes. Suitable relationships have not yet been determined for estimating conditions for mixed grain sizes, nor for intense, localized turbulence, such as eddies.

Under turbulent or fully rough flows for $R_e \geq 100$, F_* assumes a constant value of about 0.06. Thus, knowing γ_s and d, the critical shear stress can be estimated. If the estimated tractive force $\bar{\tau}_o$ from eq 52 exceeds $\tau_{o(cr)}$, particles at the bed are assumed to be in motion. Under

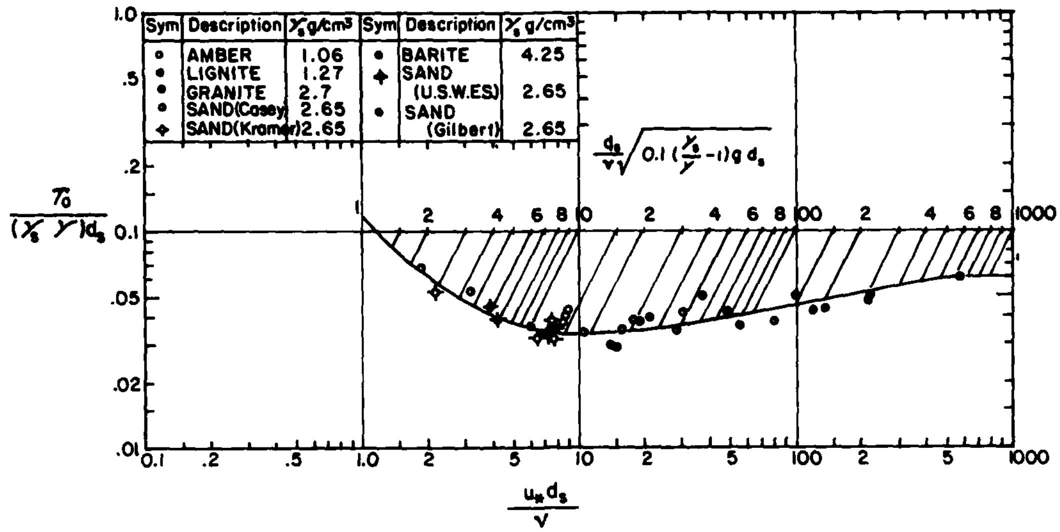


Figure 57. Shields diagram for calculating critical shear stress (after Shields 1936).

smooth or laminar flow conditions for $R_e \leq 2$, the entrainment function increases in value. For particles below a certain critical size (approximately 0.3 mm), an inverse relationship between critical shear stress and grain size exists and the required tractive force actually increases.

Under turbulent flow, there exists a thin, viscous laminar sublayer at the boundary of the bed and fluid. Thus below a certain grain size, particles are fully immersed in this sublayer, whereas above this size, they protrude above it and are acted upon by the turbulent flow. This hydrodynamically smooth boundary is often cited (e.g. Carson 1971) as the reason why additional tractive force is required to initiate motion. Allen (1982), however, has suggested the differences may be a function of the lift forces, which act downward under smooth or laminar flow conditions, augmenting particle weight and increasing the required boundary shear stress. Under turbulent flow conditions, lift forces act upward away from the bed. Experiments by Watters and Rao (1971) indicated that seepage issuing from the channel inhibits motion of particles in the bed by drag forces but enhances its movement by lift forces. Conversely seepage into a bed inhibits the effects of lift forces, but increases drag effects which reflect the modification of the sublayer by seepage.

Estimates of the amount of sediment moved by a particular flow are difficult to make, with problems inherent in each of the theoretical treatments typically used (Yalin 1977). One approach suggested by Foster and Meyer (1972) as being applicable to defining transport rates is that of Yalin (1963).

Yalin's (1963) equation is one derived for estimating bedload transport of cohesionless particles. It assumes that sediment motion begins when the lift force of flow exceeds a critical value and lifts the particle from the bed, at which point the drag force exerted by the flow moves the particle downstream. Particles may reenter the bed material once gravity causes it to settle back to the bed.

As used by Foster and Meyer, Yalin's equation is

$$\frac{W_s}{\rho_s \rho_w d V_* g} = 0.635 \; \delta \left[1 - \frac{1}{\sigma} \ln (1 + \sigma)\right] \tag{58}$$

where

$$\sigma = A \cdot \delta \tag{59}$$

$$\delta = \frac{Y}{Y_{cr}} - 1 \tag{60}$$

with $\delta = 0$ when $Y < Y_{cr}$

$$A = 2.45 \left(\rho_s\right)^{-0.4} \left(Y_{cr}\right)^{0.5} \tag{61}$$

$$Y = \frac{V_*^2}{(\rho_s - 1.0) \; gd} \tag{62}$$

and

$$V_* = \sqrt{g R S_\ell} \quad . \tag{63}$$

Parameters are

- R = hydraulic radius
- S_ℓ = slope of the energy gradient
- ρ_s = particle density
- ρ_w = fluid density
- g = acceleration of gravity
- d = particle diameter
- Y_{cr} = critical lift force
- V_* = shear velocity
- W_s = rate of transport expressed as a weight per unit time and the flow width
- δ = a dimensionless parameter of the excess of tractive force.

The number of particles in motion at any time is a linear function of δ. Which particles are transported is determined by particle density and diameter and Y_{cr}, which is calculated from the Shields diagram.

Foster and Meyer (1972) modified the equation for beds of multiple sizes, with the transport rate W_{si} of each particle size:

$$W_{si} = \left(P_e\right)_i \rho_s \rho_w g \; d \; V_* \tag{64}$$

where

- P = the left side of eq 58
- P_e = the effective P for particle size i in a mixture
- P_i = P calculated for a uniform material of size i.

Derivation is discussed by Foster and Meyer (1972). Its applicability is limited to cohensionless bed movement and to aggregates of cohesive particles. This equation predicted total transport rate and particle size distribution of the material in transport reasonably well for field plot data;

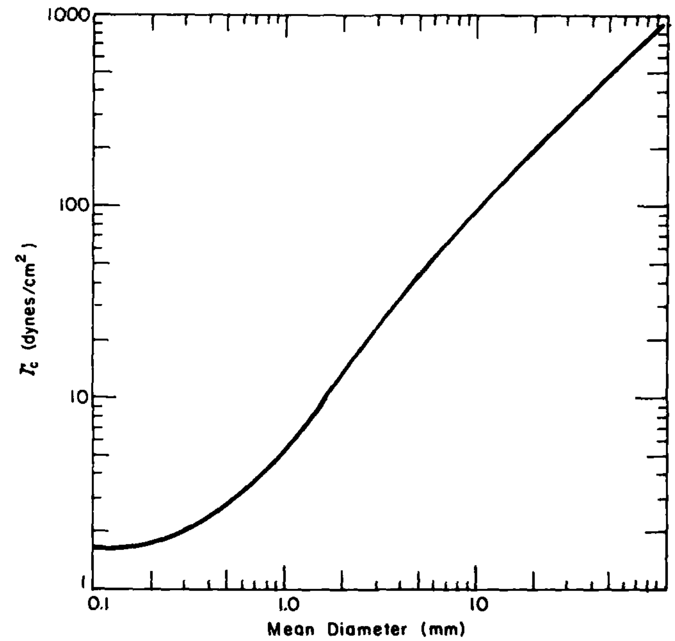


Figure 58. Critical bottom shear stress for initiation of movement of quartz sand on a plane bed as calculated from Figure 57 (after Blatt et al. 1972).

however, sediment load still differed by up to a factor of two, once rilling developed.

The relationship between grain size and critical bottom shear stress for initiation of movement of sand and gravel particles on a plane bed as calculated from the Shields relationship is shown in Figure 58.

Hjulström (1935) and Sündborg (1956) considered the initiation of movement of uniformly sized material in clear water in terms of a critical flow velocity (Fig. 59). Material in suspension increases water viscosity and hence sediment transport (Sündborg 1967) (Fig. 60). This relationship, as well as that using shear stress (Fig. 58), reveal that fine and medium size sand particles are the most easily eroded, while velocity and bottom shear stress must increase to erode progressively larger particles. In both instances, the initiation of movement of particles smaller than sand size is not well-defined by either theoretical or experimental results. In this regard, Sündbord (1956) concluded that while cohesive fine-grained particles required higher velocities than sand-size particles, those lacking in cohesion did not.

A satisfactory unique criterion for the threshold of motion of cohesive beds has not yet been devised (Allen 1982), in part because of the complexity of factors affecting erodibility and a lack of understanding of how the materials actually erode. Laboratory experiments and field observations suggest that the yield stress of the material in relation to the applied shear stress is very important (Flaxman 1963, Partheniades 1965, Lambermont and Lebon 1978); as an estimate, yield stress could be approxi-

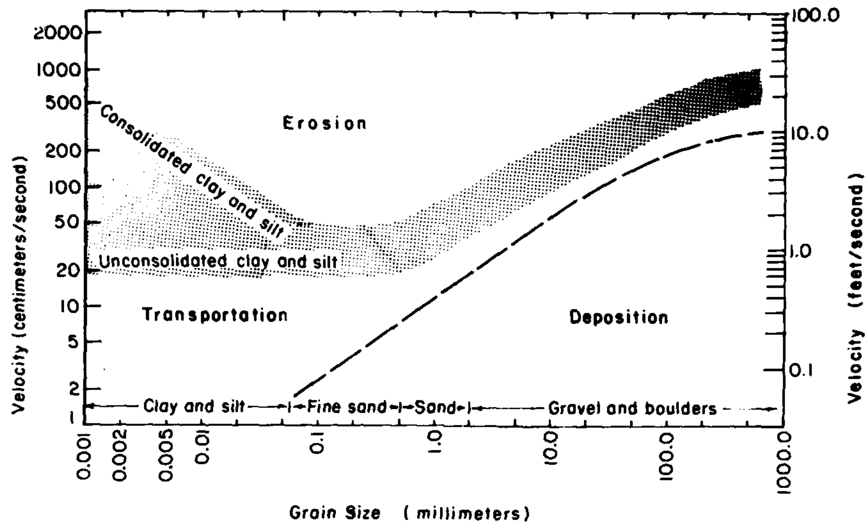


Figure 59. Critical flow velocity for initiation of motion of particles of different size as determined by Hjulström (1935). Flow velocities at which transportation or deposition will take place after mobilization are also shown. Critical velocities for erosion of cohesive silt and clay are not well defined.

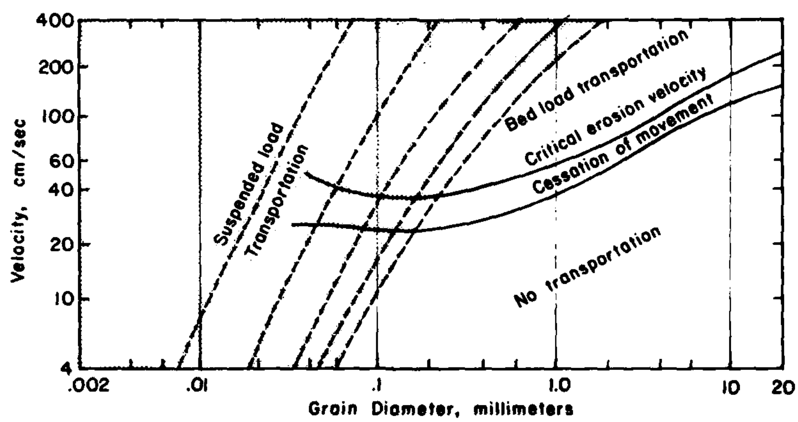


Figure 60. Effects of suspended material on particle movement (after Sundborg 1967).

mated as the inverse of the water content. Cohesive bed material generally erodes at an increasing rate as the applied fluid stress and velocity at the sediment-fluid interface increases (e.g. Partheniades 1971, Grissinger 1966, Ariathurai and Arulanandan 1978, Kamphuis and Hall 1983). These authors stressed that the cohesive material's properties at the moment of failure are important in determining the required shear stress. Lambermont and Lebon (1978) related the shear stress required to cause erosion to the yield stress of the cohesive material (τ_y) by

$$\tau = \rho_v \, p^2 \, \tau_y^{\,2q} \tag{65}$$

where ρ_v is density of fluid in a boundary layer at the interface and p and q are constants related to density of sediment at the interface.

Observations and experiments have shown, however, that a number of factors affect the erodibility or threshold stress of cohesive materials. These include grain size, clay mineral content and type, degree of consolidation, depositional history, organic matter content, physico-chemical forces related to pore fluid composition and interparticle bonds, water content, temperature, and weathering history, among others (e.g. Smerdon and Beasley 1959, Dunn 1959, Grissinger 1966, Postma 1967, Partheniades and Paaswell 1970, Partheniades 1971, Einsele et al. 1974, Ariathurai and Arulanandan 1978, Ariathurai and Krone 1976, Grissinger et al. 1981, Grissinger 1982, Kamphuis and Hall 1983).

Additionally, mud beds may erode as aggregates, rather than individual particles, thus limiting the application of a relationship between individual particle sizes and stress or velocity. Experiments by Kamphuis (1983) have also indicated that water with sand entrained in it causes erosion of a cohesive bed at lower shear stresses or fluid velocities than if the water were clear. Furthermore, erosion occurred through a combination of "sand blasting" by sand particles in saltation and by "milling" by sand moving as a layer on the bed, each of which caused localized areas of erosion that coalesced to result in the erosion of the bed as a whole.

In all cases involving calculations of either a tractive force, critical velocity or critical shear stress to initiate particle motion, estimates can vary from the actual field data by one or two orders of magnitude. The relationships presented above do not account for variations inherent in the properties of sediments, or of flow dynamics (whether sheet, rill or gully flow) under natural conditions. Overland flows are generally much shallower, intermittent and thus affected probably to a larger degree by bed roughness, and the hydraulic forces under such conditions are very difficult to measure. These factors probably account for most of the error between field data and theoretical calculations.

Overland flow factors

It is evident from research to date that numerous factors may interact to control or modify overland flow processes and determine the locations and rates of slope erosion. They may either increase the resistance of slope sediments to erosion or increase the erosiveness of these processes.

Erosion resulting from overland flow processes, as a group, are usually considered a function of several key factors. These factors affect the magnitude of erosion by affecting either raindrop impact or erosive action of runoff (e.g. Smith and Wischmeier 1957, Young and Mutchler 1969b, Culling 1963, Emmett 1970):

1. Sediment type and properties. The infiltration capacity of the sediment composing a slope and its resistance to the forces of rainfall and flowing water are crucial. As is clear from the previous discussion, grouping materials as cohesive or noncohesive/cohesionless is an important distinction related to shearing resistance and yield stress. Properties of both these sediment groups are related to their sedimentary origin, which largely determines grain size, grain shape, particle packing and weathering

history. Weathering may result from both long- and short-term processes affecting sediment strength and structure, such as leaching, moisture changes, desiccation and consolidation, and those affecting infiltration capacity, such as compaction, composition and presence of fines. Clay content and mineralogy are important in this regard. The frozen state effectively increases resistance to erosion, but thawing or cycles of freezing and thawing may produce weakened material more susceptible to failure or scour. As with other erosional processes, complex slopes offer differing resistances to erosion related to their stratigraphy, structure and composition.

2. Slope geometry. As the angle of slope increases, the velocity and hence erosiveness of runoff increases, and sediment and slope stability decrease. Splash erosion apparently increases, mainly due to the increasing influence of the acceleration due to gravity, which outweighs the reduced effect caused by the angular impact of drops on slope particles. Height of slope is important for the same reason. For example, a doubling in velocity increases kinetic energy fourfold which in turn increases the volume of sediment that can be transported by 64 times (Zachar 1982). An indirect effect of increasing transport capacity is to decrease surface roughness with respect to flow. In contrast, lowered slope angles may induce deposition of eroded material.

Slope length principally affects flow depth and confluence, but its effect is not necessarily to increase erosion, since the carrying capacity of runoff may decrease or be exceeded, if with movement further downslope, the turbulence or volume decreases. Slope length and inclination are related in this respect. Slopes which change in steepness over varying lengths combine conditions conducive to erosion and sedimentation.

Slope aspect principally affects net solar radiation input and thus soil moisture and its variability, and soil temperature, which affects rates of thawing or freezing, decomposition, and wetting or drying. Orientation with respect to primary wind directions and hence raindrop impact will affect rainsplash.

3. Vegetation. The percentage cover, density and root structure of vegetation determines protection of slopes to erosional processes. For example, vegetation protects against raindrops, increases the degree of infiltration of water by obstructing water flow and reducing its velocity, mechanically binds soil, and diminishes near-surface fluctuations in microclimatic variables. Soil development inherent with vegetation growth generally improves soil cohesion and possibly infiltration capacity. Absence of vegetation has been shown to result in a one- to two-fold difference in the magnitude of erosion losses.

4. Climate. Climatic effects include temperature, solar radiation, precipitation and their seasonal variability. Storm activity (intensity and frequency) clearly influences the erosiveness of rainfall, sheet wash and rill flow, which in relation to snow cover and frozen ground, may likewise increase or diminish erosiveness.

Erosion predictions

Within the field of agricultural engineering, research has focused on developing an equation to predict soil erosion rates by rainfall, sheet wash and rill erosion on agricultural lands in differing climatic, geographic and physiographic conditions. One result of this research is the Universal Soil-Loss Equation (USLE) as proposed by Wischmeier and Smith (1965). It is empirically based upon the statistical analysis of measurements from across the United States. As such, it is established for average soil losses from agricultural lands and thus generalizes and lumps many of the complex and poorly understood erosional factors and processes (e.g. Morgan 1983).

The USLE is

$$A = R\ K\ L\ S\ C\ P \tag{66}$$

where

A = the computed soil loss per unit area
R = the rainfall factor
K = the soil-erodibility factor
L = the slope length factor
S = the slope-steepness factor
C = the cropping management factor
P = the erosion-control practice factor.

Each factor has been calculated for use in specific geographic localities in Wischmeier and Smith (1965). A brief description of what each factor accounts for is given below.

The rainfall factor R is a measure of rainfall erosion and runoff that is based upon rainfall intensity, duration and frequency, and calculated from the kinetic energy of rain over 30 minutes.

A measure of the susceptibility of a soil to erosion is given by K. It accounts for the infiltration capacity and resistance of the soil to detachment and transport by rainfall and runoff.

The effects of slope length on the increase in downslope runoff on soil loss is measured by L. Similarly, S considers the effects of increased slope angle on runoff velocity and is based upon a quadratic equation defined by empirical data. Both factors are normally combined and a slope factor chart is used to determine LS (the factor of soil-loss).

Both C and P account for land use practices such as the type of crop rotation or tillage methods, or the conservation practices such as contouring or strip-cropping that are followed. Again, tables of values have been determined.

In use, farmers or land-use planners would determine a tolerable level of soil loss, set that equal to A and then determine practices for C and P for their conditions as considered by R, K and LS. For natural slopes without vegetation both C and P would equal 1.

Zachar (1982) indicated that soil loss A_s due to snowmelt runoff could be estimated with the following change to the USLE:

$$A_s = mhk\ LSCPK \tag{67}$$

where m is the rate of thawing of snow over a 20-day period, a value which is 50 to 100% larger in regions where rain enhances snowmelt, k a rain-water runoff value, and h the amount of water derived from snowmelt over the same 20-day period. Further modifications would also be needed if the snow cover were discontinuous or in drifts.

Application of this equation or the USLE to erosion along shorelines is basically untested. Sterrett (1980) applied a modified form of the USLE to steep, eroding bluffs along Lake Michigan. This modification was proposed by Foster and Wischmeier (1974) for evaluating irregular slopes. His field data did not closely correlate with the modified USLE prediction, but he suggested that this may be the result of inaccuracies in field measurement techniques as well as the applicability of the equation, which typically overestimated the volume of sediment lost. No account was made of changing conditions during snowmelt runoff.

A mathematical model deserving analysis for shore erosion was proposed by Meyer and Wischmeier (1969) to simulate the processes of slope erosion due to particle detachment and movement by raindrops and erosion, and to transport by runoff processes. The model permits variation of the steepness, length and shape of the slope, rainfall intensity, soil infiltration rate, and factors affecting soil erodibility in order to estimate net erosion. The model, however, remains in need of additional calibration by using field data to enhance its applicability.

GROUND WATER EROSION

Ground water is important as an agent of erosion and in modifying the geotechnical properties of sediments to increase their erodibility. The importance of water in decreasing the stability of slope sediments through the development of excess pore pressures and seepage pressures and changes in soil moisture have previously been discussed. Antecedent soil moisture conditions are also clearly important in determining the susceptibility of materials to erosion (Wolman 1959, Hooke 1979). In this section, I discuss ground water erosion caused by piping and the discharge of springs that are related to complex patterns in the ground water flow system and geology.

Both the regional and shore zone geology can complicate the water table configuration and direction of ground water flow adjacent to reservoirs. Under natural conditions, small valleys or lakes may act as local or regional ground water recharge or discharge locations, or a through-flow condition may exist (Fig. 61). On regionally sloping terrain, through-flow results from inflow of ground water into the upslope end of the basin and outflow of ground water at the downslope end (e.g. Sloan 1972, Born et al. 1979). Most reservoir pools located in former river valleys apparently act as regional discharge locations, but ground water flow patterns around artificial impoundments have received little study.

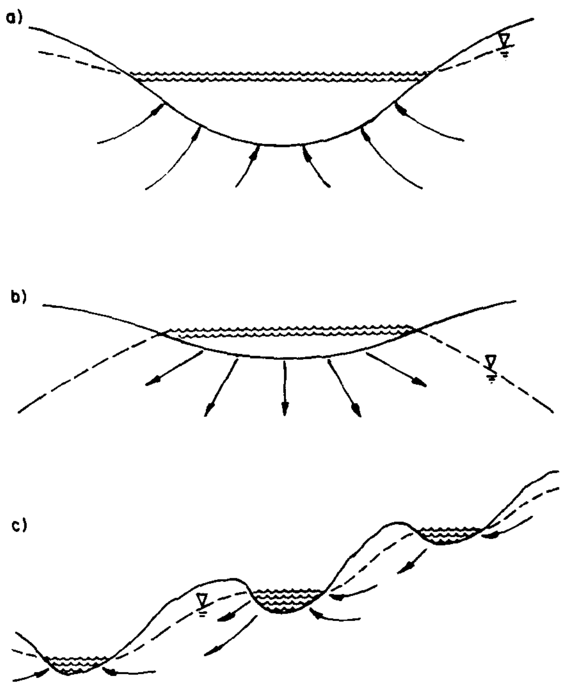


Figure 61. Ground water flow systems for lakes or valleys acting as discharge (a) or recharge (b) locations, or with a through-flow condition (c).

On a more localized scale, the flow pattern and water table configuration within particular reaches of a reservoir's shore zone may be more directly influenced or controlled by the local geology, reservoir pool operations, man's use of ground water reserves (e.g. wells), and seasonal variability in ground water recharge (e.g. snowmelt [Meyboom 1966]). The general interaction of water levels and water table fluctuations was discussed in the section Water Level Fluctuations.

Variability in the physical properties of sediments and in their stratigraphic relationships can complicate the ground water flow system, with minor geological details possibly playing a major role in determining seepage patterns (Terzaghi 1929). With homogeneous shore zone sediments, flow patterns are determined largely by the regional ground water flow pattern and hydraulic gradient and movement of the impoundment's water level. In contrast, the hydraulic conductivity is more important in shore zones composed of inhomogeneous sediments; less pervious or impervious beds may result in multiple transient or permanent perched water tables. Glacial materials are generally inhomogeneous and especially characterized by complex flow systems (e.g. Knuttson 1971, Grisak and Cherry 1975, Born et al. 1979, Sterrett and Edil 1982).

Multiple layers of alternating impervious and pervious sediments can result in multiple piezometric surfaces, each related to a particular stratigraphic unit (Fig. 62). Discontinuous layers or lenses of coarse material will be characterized by higher flow rates and piezometric surfaces. Even multiple thin layers of millimeter thickness, such as in the varved sequences described by Deere (1957) and Deere and Peck (1958), can be characterized by higher and lower discharges at slope faces. Fractures or joints may provide avenues of flow in otherwise dense material of low permeability; for example, in a clay-rich till, the flow within the joints can

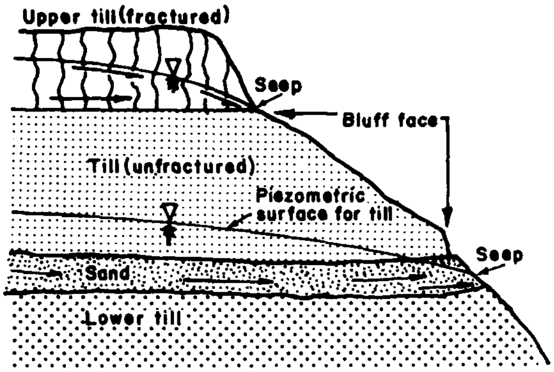


Figure 62. Ground water flow systems in a bluff composed of glacial deposits on the shore of Lake Michigan (after Sterrett and Edil 1982).

result in effective permeabilities for the unit that are much higher than that for the material itself (e.g. Williams and Farvolden 1967, Grisak and Cherry 1975, Lafleur and LeFebrve 1980, Hendry 1982). An example of a complex ground water flow system in glacial deposits composing bluff's along Lake Michigan is described by Sterrett and Edil (1982, Fig. 62).

Some situations of complex ground water flow systems that result in ground water erosion include:

1. A water level rise in the impoundment followed by a drawdown sufficiently rapid that seepage from shore materials cannot maintain the water table coincident with the pool level. Springs or seeps may develop within the pervious strata of bluff or beach faces. Their effects on stability were previously described.

2. In a bluff of pervious materials overlying relatively impervious materials, water levels may rise and be maintained over the short term by an increase in local recharge from snowmelt runoff or heavy precipitation. Again, seepage of significant discharge and instability can result.

3. In a bluff of interstratified coarse- and fine-grained sediments, a rise in pool level can dump water into a pervious layer that is high in the stratigraphic section and above impervious strata. Pool lowering results in perching of the water and seepage at the face.

4. Thawing of seasonally frozen soil at the ground surface releases ice in the ground as meltwater, which saturates and then flows within the thin surface layer above the relatively impermeable frozen soil.

5. Frozen ground, whether seasonally or perennially frozen, can form an impermeable barrier that restricts or blocks ground water flow below it, thereby permitting a buildup in head and pressures at depth that can be released by rapid outflow during a winter thaw or as a seep during the spring thaw (e.g. Kane et al. 1973).

The importance of perched or raised water levels, relative to the pool level, is that the discharge of ground water can physically erode bluff, and sometimes beach zone sediments, where they emerge. Outflow may be sufficient to actually detach particles or aggregates and transport them into

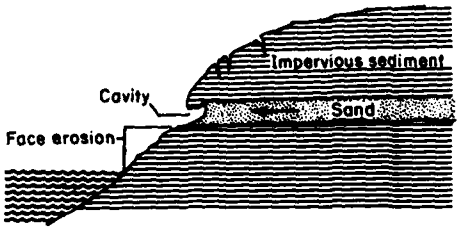


Figure 63. Discharge of ground water as springs or seeps from the pervious sand may cause removal of sediment at the point of discharge as well as internal erosion or piping of fine-grained material within the slope. Sheet and rill flow erosion may take place below the point of discharge as well.

Figure 64. Bluff face degradation resulting from discharge of ground water from sand lens in sequence of glacial deposits in Iowa.

the impoundment (e.g. Deere 1957, Weigel and Hagerty 1983) (Fig. 63). This detachment occurs if the seepage force is sufficient to overcome the sediment's resistive forces, which mainly depend upon internal cohesion, particle interlocking and weight (e.g. Sherard et al. 1963).

Degradation of bluff faces can progress backward into the permeable unit, creating a conduit or cavity (e.g. Deere 1957, Sherard et al. 1963, Palmer 1973, Hadley 1976) (Fig. 64). This progression in turn removes sup-

Figure 65. Collapse of top of bluff composed of glacial deposits by piping and subsurface erosion of sand lens.

port under portions of the slope and failure of the overlying sediments may follow (Fig. 65). Water emerging from stratigraphic units that are in the upper parts of slopes may also form into sheet and rill flows, eroding bluff or beach sediments beneath the seepage point (e.g. Deere and Peck 1958, Hill 1973, Weigel and Hagerty 1983). Fine-grained sediment beneath seeps may also become sufficiently saturated to fail and flow (e.g. Edil and Vallejo 1977).

Internal or subsurface erosion of permeable layers, or piping, may also take place. Fine-grained sands or silty sands are more susceptible to this process than coarser materials. Dirty water emerging from an otherwise non-eroding face is a clue that piping is occurring. As water moves through the sediment, the pressure head is gradually dissipated, but concentrated flows can also generate sufficient seepage force to move or entrain individual particles (Casagrande 1936). Reduction or loss of particle interlocking and collapse of the permeable bed may then result in liquefaction of the layer and failure of the otherwise stable, overlying bluff sediments (Casagrande 1936, Terzaghi and Peck 1967). Actual conduits or cavities beneath cohesive or peaty units may develop by piping (e.g. Casagrande 1936, Jones and Crane 1981), although this erosion is apparently less important than degradation and erosion caused by seeps in the bluff face.

Similarly, concentrated flow within joints or fractures may preferentially remove fine-grained materials or leach out calcite and iron oxides, thereby locally reducing the resistance to shearing and developing a failure condition along these anisotropies (e.g. Deere 1957, Sterrett and Edil 1982).

If waters emerge from submerged or saturated sediment under a head, whether in slope or not, sufficient uplift force may be generated to exceed the gravitational force and heave them. Boils may develop in a situation analogous to heave resulting from piping along the base of a dam (Terzaghi and Peck 1948).

A contrasting effect occurs during winter when seeps or springs may develop icings on bluff faces and beach sediments (e.g. Carey 1973). These icings will protect shore zone materials from waves, currents or other subaerial erosive forces during winter and early spring, but during thaw, the meltwater may saturate underlying sediments and create a stability or erosional problem.

The mechanics of piping can be considered in terms of forces on individual grains within the permeable sediment or soil. The force acting on a grain, known as the seepage force F_s, is defined in terms of the hydraulic gradient and calculated by

$$F_s = \gamma_w \; g \; \Delta h \; A \tag{68}$$

where

- A = the cross-sectional area of the grain
- γ_w = specific gravity of water
- Δh = the difference in the hydraulic head between the front and rear faces of the particle (e.g. Cedergren 1977).

In terms of flow within a particular volume of soil, this simplifies to

$$F_s = \gamma_w \; g \; i \tag{69}$$

and seepage force is directly proportional to i. Terzaghi and Peck (1967) define i γ_w as the seepage pressure.

Thus in areas of downward percolating ground water, seepage forces act in the direction of gravity; however, in areas where water is moving upward, seepage force opposes gravity, and particles can be entrained and removed if the seepage force exceeds their effective weight. Calculations are generally made by using flow nets; this technique is described in detail in Cedergren (1977).

THERMAL CONDITIONS

Whether shore zone materials are frozen, unfrozen or in transition between these two states can significantly affect the rate, style and occurrence of erosion. The influence of frozen ground and of repetitive freezing and thawing has been recognized as affecting stability and erodibility for some time, and more recently their overall importance to shore or bank

erosion and recession has been verified (e.g. Chieruzzi and Baker 1958, Wolman 1959, Owens and McCann 1970, Hill 1973, Sterrett 1980, Reid 1984). Seasonal variability in erosion and recession rates may result from seasonal changes in ground thermal conditions. Several authors (e.g. Wolman 1959, Hill 1973, Sterrett 1980) concluded that over 50% and up to 96% of bluff zone erosion took place in winter or early spring in response to freeze-thaw and related processes.

Seasonal shifts in air temperature of northern regions produce similar shifts in near-surface ground temperatures that can be simplified, for purposes of discussion, as conforming to four temperature periods: an unfrozen summer period, a frozen winter period, and fall and spring seasons characterized by either a gradual freeze-up or gradual thaw that is punctuated by diurnal or short-term periods of thaw and refreezing or freezing and re-thawing, respectively. In arctic regions where ground at depth is perennially frozen, such seasonal shifts in ground temperature are basically limited to a thin active layer at the ground surface. A brief discussion of shore zones with perennially frozen materials in them will complete this section.

The rate, timing and depth of freeze-up and thaw are determined by numerous factors that determine the thermal properties of soils such as soil composition, structure, density, porosity, water content, degree of saturation and temperature. Other factors such as slope aspect, vegetative cover, snow depth and air temperature are interrelated with these soil properties in determining soil freezing or thawing. Farouki (1981) recently reviewed the literature on the thermal properties of soils and the factors determining thermal conductivity.

In general, freeze-thaw cycles common to the spring and fall affect soil (sediment) properties such as the degree of consolidation, structure, permeability, strength, moisture content and density (e.g. Slater and Hopp 1949, 1951, Slater 1951, Willis 1955, Sillanpää and Webber 1961, Broms and Yao 1964, Culley 1971, Chamberlain and Blouin 1978, Chamberlain and Gow 1979, Johnson et al. 1979). The most significant changes take place when water is readily available at the ground surface or just below it. Physical changes may result from either the in-situ freezing of water in the sediment and expansion ($\approx$ 9%) accompanying this change of state, or from freezing and the formation of ice lenses when water migrates from within underlying sediment or from the underlying water table toward the downward advancing freezing front (e.g. Taber 1929, 1930, Beskow 1935, Penner 1959, Takagi 1965, 1979, Miller 1972).

The ice that freezes within the soil or sediment may take several forms and produce different textures that vary in porosity and permeability. Dingman (1975) found that four textures are generally recognized: 1) "concrete frost," essentially saturated or supersaturated ground that is completely frozen (save for negligible amounts of unfrozen films on soil grains), and which may contain ice lenses; 2) "granular frost," with small ice crystals intermixed with soil particles and aggregated around them; 3) "honeycomb frost," which is similar to granular, but with a higher degree of connection among ice crystals and a lower porosity; and 4) "stalactite frost," apparently identical to "needle ice" or "piprake," which is characterized by small, needle-like ice crystals aligned vertically and extending

downward into the soil from a heaved surface. Concrete frost is generally very low in porosity and often considered impervious. It is thus similar to bonded perennially frozen ground that is supersaturated and ice-rich. In contrast, granular frost and stalactite frost are relatively permeable and absorb moisture, and can prevent overland flow of runoff although still frozen (e.g. Haupt 1967).

The segregation of ice in the form of lenses or crystals most often leads to a lifting or heaving of the ground surface. The amount of heave is mainly dependent upon the dimensions and directions of growth of the ice lenses or crystals (typically of 1 millimeter to tens of centimeters thick), but also the compressibility of the sediment (e.g. Chamberlain and Blouin 1978, Chamberlain and Gow 1979). Spatial dimensions of ice lenses appear to depend upon the frequency of freeze-thaw cycles, the temperature gradient or thermal conductivity of the soil, the availability of water near the freezing front, and the ability of the soil to transmit water.

The mechanics of ice segregation and frost heave remain without a definitive theory although three fundamentally different theories (capillary, secondary heave and adsorption force) are the primary subjects of research and considerable discussion (Chamberlain 1981). While different in approach, properties that affect the susceptibility of sediments to frost heave are similar in each theory.

Chamberlain (1981) concluded that principally seven factors are important in determining a soil's susceptibility. These factors are the soil's texture (grain size, gradation and mineralogy), pore size, moisture content of the sediment, rate of heat removal during freezing, temperature gradient, overburden stress or surcharge, and the number and duration of freeze-thaw cycles. How they affect frost action remains uncertain, with research providing some contradictory evidence of their increasing or decreasing susceptibility (Chamberlain 1981). Tests to determine a soil or sediment's susceptibility are not standardized, apparently because of a lack of understanding of these and other factors that control frost heave; a detailed description of the over 100 tests used today is given in Chamberlain (1981).

Even so, grain size is typically used as the indicator of frost-susceptible soils, with those containing some fraction of finer grained particles assumed to be particularly affected by it. Quite often, the limits of Casagrande (1931) -- non-uniform soils consisting of more than 3% of their particles finer than 0.02 mm in diameter and uniform soils of more than 10% of their particles smaller than 0.02 mm diameter -- are used to denote frost-susceptible materials. Silt-rich soils are particularly susceptible, but clay-rich are much less so because of their low permeability. Because of the inherent variability of properties and conditions of natural slopes along reservoirs, frost susceptibility and hence the effects of freeze-thaw and ice segregation are likely to vary from site to site, and at a site, within individual units of bluffs composed of complex sedimentary sequences.

<u>Freezing and thawing effects</u>

As previously stated repetitive freezing and thawing, with or without ice lens formation, can significantly modify physical properties of surfi-

cial materials. Changes that clearly increase the erodibility of these materials include substantial reductions in shear strength and bearing capacity accompanied by a decrease in compaction, density, and degree of consolidation and an increase in porosity and permeability (Tsytovich et al. 1959, Farouki 1981). Broms and Yao (1964), for example, found that cyclic freezing and thawing of clay-rich sediment reduced unfrozen shear strength by up to 95%, with the greatest reduction occurring in sediment that had the highest moisture content before freezing. In addition, the factors modified by frost action affect thermal conductivity and thus soil freezing or thawing (Farouki 1981).

The repetition of freezing and thawing with ice lens or crystal formation will crack, disaggregate, separate and reorient soil particles or aggregates while modifying their structure (e.g. Willis 1955, Logsdail and Webber 1959, Sillanpää and Webber 1961, Tsytovich et al. 1959, McDowall 1960, Schumm and Lusby 1963). Soils with a significant clay fraction undergo shrinkage and compression as water is removed to form ice lenses (Tsytovich et al. 1959). Similarly, freezing cohesive soils may cause agglomeration of the fine-grained materials, with small fractures forming between the aggregates and filling with ice (Czeratzki and Frese 1958). A layering of ice lenses with their spacing increasing with depth may also develop, apparently when a slow freezing rate is accompanied by balanced heat and moisture flow at the freezing front (Palmer 1967). Preferential ice lens formation within fractures, joints or bedding planes may widen these features and weaken the materials along them; this can create an unstable situation with failure surfaces located on these features (e.g. Hill 1973).

Active erosion by soil creep on sloping surfaces results from heaving of soil particles by ice formation. Heaving results from the vertically upward movement of particles caused by soil expansion, which is then followed by thawing and the downslope displacement of the particles as they settle under the influence of gravity (Davison 1889, Washburn 1967, 1969, Benedict 1970, 1976). Expansion or heaving is proportional to the total thickness of segregated ice crystals and is generally in a direction at right angles to the ground surface. Vegetation can severely limit the efficiency and importance of creep due to frost action (Schumm and Lusby 1963). On steep slopes exposed to sun in winter, particles and aggregates lifted and separated by ice formation may be released by sublimation of the ice, allowing them to fall or roll to the slope's base where they remain until acted upon by waves or currents in the spring (e.g. Wolman 1959, Harrison 1970). Wind may also be more effective on such loosened materials. Spring thaw and ice melt may similarly release heaved particles.

On lower angle slopes of bluffs or beaches, heaving and thawing of the ice under unsaturated and unconfined conditions produces loose, weakened and fluffed surficial materials that are more susceptible to erosion by tractive forces exerted by waves, overland flow or the wind (e.g. Chieruzzi and Baker 1958). However, the increase in permeability and infiltration capacity may actually limit sheet and rill flow of meltwater and precipitation (e.g. Schumm and Lusby 1963, Haupt 1967).

Seasonal thawing

The seasonal thawing of frozen sediment releases meltwater from pore and lens ice that can significantly reduce internal friction and cohesion and thus decrease shearing resistance (e.g. Williams 1959, Nixon and Hanna 1979). If the underlying still-frozen sediments are bonded and essentially impervious, excess pore pressures may develop at the ice/sediment interface that can reduce or eliminate shearing resistance of this material (e.g. McRoberts and Morgenstern 1974a,b, Nixon 1973). In general, the frozen horizon prevents free drainage of water through the soil and maintains saturation of the uppermost thawed material, thus encouraging downslope movement (Benedict 1976). Gravitational slip or flow type failures on slopes as low as 1° to 10° may result (e.g. Atkinson and Bay 1940, Tigerman and Rosa 1949, Sterrett 1980). Additionally the migration of meltwater produces seepage pressures that likewise reduce shear strengths and enhance erodibility.

Because thaw proceeds from the top and bottom of the frozen layer, subsurface drainage of meltwater may be impeded and hydrostatic pressures built that, upon release, can result in a rapid outflow as a seep which can cause piping. The basal thawed layer may be susceptible to liquefaction in this state (McRoberts and Morgenstern 1975, Finn et al. 1978).

On beaches, thaw progresses approximately parallel to the beach face, generally deepening through the spring (e.g. Harper et al. 1978). The depth to remaining ice-bonded beach sediments limits the effectiveness of waves and ice to cause erosion (Owens and McCann 1970, Harper et al. 1978); however, intense wave activity during storms can increase thaw rates and thermally or mechanically erode blocks of material from the frozen beaches (Taylor 1980, 1981).

The seasonally frozen condition

A seasonally frozen condition affects erosion and recession in five major ways:

1. Frozen ground that is bonded by concrete ice offers significantly more resistance to erosional forces such as waves, currents, rainfall and runoff (e.g. Schumm and Lusby 1963, Haupt 1967, Miles 1976, Scott 1978), and, at least for cohesionless sediment, is generally more stable than when unfrozen (e.g. Eardley 1938, Gill 1972, Walker 1969, Taylor 1980). Normal beach and nearshore processes are slowed or stopped after beach freeze-up (e.g. Davis et al. 1976).

2. Inundation of frozen, coarse-grained littoral zone sediments will fill pores with water that soon freezes (Abele 1977). This effect can produce a fully ice-bonded gravel or very coarse sand within which segregation ice would not normally develop. Such materials will resist erosion by waves, but observations by Taylor (1980) suggest that gravels will erode particle-by-particle as ice melts. In interstratified beach or bluff materials, this can undermine and cause erosion of still-frozen blocks of finer-grained sediment (Taylor 1980).

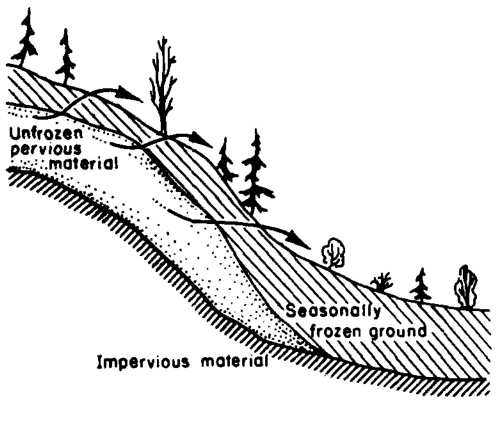


Figure 66. Possible situation on a slope in which frozen ground acts to confine ground water flow, increase hydrostatic head and force water to the ground surface in winter, where it would freeze to form aufeis (after Carey 1973). A similar situation is possible along reservoir or lake shores.

3. The presence of ice-bonded sediment (concrete frost) prevents infiltration and increases the rate of surface runoff. However, when frozen ground with lower ice contents and granular, stalactite or honeycomb frost exists, the permeability and infiltration capacity may actually be greater during winter after frost heaving has taken place (e.g. Schumm and Lusby 1963, Haupt 1967, Soons and Rainer 1968, Dingman 1975). Seasonally frozen sediment with a low ice content generally behaves similarly to unfrozen sediment and can have high infiltration rates (e.g. Kane and Stein 1983).

4. The formation and continued existence of a frozen surface layer may form an impervious barrier that prevents the flow of ground water from the base of slopes and leads to an excess hydrostatic head (Siple 1952). This pore pressure rise reduces the resistance to shear within the unfrozen saturated zone, as well as increasing its weight, and can result in failure and slumping of the bluff. Slab failures may likewise take place (Sterrett 1980). Without a slope failure, sufficient head could develop and cause water to burst through the frozen cover or be forced to flow along anisotropies within it, with possible piping and aufeis formation (e.g. Stieglitz 1978) (Fig. 66). The presence of a relatively impervious base is required to prevent dissipation of the hydrostatic head.

5. The boundaries of large ice bodies, or perhaps the trends of surfaces of relatively horizontal ice lenses, can act as planes of weakness (e.g. MacCarthy 1953, Walker and Arnborg 1966). Thawing on the ice surface increases the possibility that this discontinuity will become a potential failure surface (Williams 1959).

The perennially frozen condition

The effects of a perennially frozen condition (permafrost) on the manner and rate of shore erosion are not yet understood. Controversy exists as to whether permafrost inhibits lateral erosion and shoreline recession or actually increases the rate and extent of erosion and recession. Unfortunately, systematic and quantitative measurements of eroding perennially frozen shores or streambanks have not been made (e.g. Cooper and Hollingshead 1973, Newbury 1974, Scott 1978, Shur et al. 1978, Lawson 1983a).

Thermal and mechanical processes appear equally important in causing recession of perennially frozen streambanks or shore zones. Are (1973, 1977) considered the erosional processes to be 1) thermoabrasion -- erosion

of shore sediments by the thermal and mechanical energy of moving water, 2) thermokarst -- subsidence resulting from thaw of frozen bed and bank materials containing excess ice, and 3) thermodenudation -- subaerial processes eroding bluff materials thawed by warm air and solar radiation. Thus the processes that affect the thermal and physical states of the material are those that determine erosion and recession rates.

Factors considered important in affecting the thermal state are the grain size, stratigraphy and ice content of the materials, extent and thickness of the vegetative cover and root or peat mat, bluff aspect and slope, thickness and duration of the ice and snow cover, water level and temperature, and climate (e.g. Nixon and McRoberts 1973, Outhet 1974a, Miles 1977, Are 1977, Harper 1978a,b, Ritchie and Walker 1974, Scott 1978).

Evidence available to date suggests that probably the most important physical factors affecting the erodibility and stability of perennially frozen shore zones are the type, content and distribution of ground ice (e.g. Are 1973, 1977, 1983, Trepetsov 1972, Tomirdnaro and Ryabchun 1974, Miles 1977, Lawson 1983a, Harry, et al. 1983). In addition, erodibility remains a function of the sediment's grain size, structure and composition and the extent of the vegetative cover (e.g. Eardley 1938, Williams 1952, Tomirdnaro and Ryabchun 1974, Serov and Leschchikov 1978, Scott 1978, Walker 1983, Newbury and McCullough 1983).

Geotechnical properties of sediments while frozen and while thawing are affected by the quantity of ground ice (Andersland and Anderson 1978, Johnston 1981) and bearing capacity, slope stability, settlement and shear strengths may differ. Substantial quantities of ice (60 to 90% by volume) have been observed in sediments at depths of up to 45 m, with ice varying from small pore fillings and veins to massive bodies in the form of lenses, layers and wedges (e.g. Black 1969, Rampton and Mackay 1971, Trepetsov 1972, MacKay 1972, Brown and Sellmann 1973, Williams and Yeend 1979, Pollard and French 1980, Lawson 1983b). The occurrence and distribution of ground ice depends on several factors that remain poorly understood and it is impossible at present to predict its presence (Mackay and Black 1973, Mackay et al. 1978). Fine-grained sediments commonly contain large quantities of ice but there are significant exceptions that are a function of geologic factors (Lawson 1983b).

Thawing of ice-rich permafrost results in subsidence as ice melts and the sediments consolidate (Muller 1947, Terzaghi 1952). Excess pore pressures may develop that can significantly reduce the strength of the thawed material (Nixon and Hanna 1979), making it highly susceptible to erosion by waves or currents, as well as reducing its stability and causing failure of the slope (e.g. Morgenstern and Nixon 1971, McRoberts and Morgenstern 1974a). Melting of massive ice can rapidly generate excess pore pressures (Nixon 1973), which in addition to modifying material strength, can reduce shear strengths on the melting ice surface and result in a slip failure along it. Slip failures also occur along the ice surface when it is a plane of weakness, typically when bluff materials are oversteepened by toe erosion (e.g. MacCarthy 1953, Walker and McCloy 1969, Harper 1978a, Church and Miles 1983). Because permafrost is generally impermeable, it restricts meltwater to the thawed layer of sediment at the ground surface and allows seepage and excess pore pressures to develop as the thaw depth increases

Figure 67. Thermoerosional niche eroded into perennially frozen shore materials by wave action. Collapsed overlying sediments which have thawed protect lower bluff sediments until eroded away.

(Williams 1970). Saturated thawed sediment can fail and flow on very low angle slopes of 1° to 15°, thereby exposing the permafrost beneath it to further thawing and degradation. Waves, currents and overland flow processes may also actively remove thawed shore zone sediment.

Are (1973, 1983) concluded that a critical ice content exists at which shore zones will continue to recede because of thermal erosion. Pure ice shores would retreat indefinitely under the simple presence of water above 0°C. In contrast, if shore zone materials contain no ice, thermal energy of the water would have no direct influence on shoreline recession. In addition, if the ice content of the shore zone sediments is sufficient to result in thaw subsidence of the ground level below that of the water level in the reservoir, shoreline retreat would again proceed unimpeded (Are 1983). The ice content typically lies between these extreme conditions. Harper (1978b) also concluded that for a given amount of available energy for erosion and sediment transport, ice effectively increases the amount of erosion that can result from that energy, since it does not contribute to the sediment load of eroding currents. Rates of retreat of ice-rich bluff sediments, particularly those with massive ground ice, are some of the higher ones reported (e.g. Mackay 1963). Thaw subsidence of submerged frozen sediments will also take place as the water moves inland, which could maintain deeper water near the reservoir shoreline and prevent beach formation.

A typical pattern of erosion of perennially frozen beaches and bluffs is as follows. Attack by waves and currents erodes a niche into the frozen

Figure 68. Thawed and frozen blocks of sediment which have failed by collapse due to undercutting by waves. Ground ice (photo at center) apparently acted as a failure plane along which some material has slid into the water.

sediment at the water's edge (e.g. Leffingwell 1919, Williams 1952, Czudek and Demek 1970, Gill 1972, Walker 1978, 1983). This niche (Fig. 67) is usually called "thermoerosional" because of its formation by both thermal erosion and mechanical processes (Abramov 1957). Niche development is mainly dependent upon the wave energy impinging upon the shore, water temperature and composition (Williams 1952, Lewellen 1965, Walker and Arnborg 1966). Storm-generated waves and currents can result in extremely rapid erosion and shoreline recession (e.g. Williams 1952, Walker and Morgan 1964, Harry et al. 1983).

Niche development in permafrost materials may vary in extent along reaches of the shore, depending upon its orientation with respect to prevailing wind direction, the effective fetch and water depth. Fluctuations in water level will increase niche height, with its shape and extent depending on the length of time the water remains at a particular level (Walker and Arnborg 1966, Lewellen 1965, 1972). Rate and depth of niching apparently depend upon the physical properties of the frozen sediments, with cohesive materials being more stable upon thawing, except when ice contents are high (e.g. McDonald and Lewis 1973, Outhet 1974a, Miles 1977, Scott 1978). After thawing, ice-rich cohesive materials apparently erode faster, and are less stable than, thawed cohesionless sands and gravels.

A critical depth of undercutting of the niche is eventually reached, causing failure of overlying bluff materials (Fig. 68). This depth is a

Figure 69. Thermal and physical erosion of 15-m-high bluff containing ice wedges. Preferential melting of ice wedges has left overhanging promontories of still-frozen silt and vegetation. Sediment flows are active in removing thawed material from the slope's base.

function of bluff height and frozen ground properties. Fine-grained silt and clay are typically undercut the deepest (5 to 12 m) while sand and gravel exhibit less undercutting (1 to 3 m) (e.g. Walker and Arnborg 1966, Miles 1977, Harper 1978a, Walker 1983).

Frozen sediment may fail by block collapse after being cantilevered out over the water surface. Preferential failure planes may appear along thawing or actively cracking ice wedges or other massive ice surfaces (MacCarthy 1953, Walker and Arnborg 1966, Harper 1978a,b) (Fig. 69). Intertwined vegetation mats or peat layers may resist tensional failure and become draped over the bluff surface, thereby protecting it (Fig. 70). Failed material that falls to the slope's base or within the shallow nearshore water provides a buffer to further attack of the bluff by waves and currents (Dylik 1969), but once removed, the collapse process is repeated (e.g. Are 1977, Newbury and McCullough 1983).

Additional failure mechanisms of bluffs include 1) the repeated failure and flow of thawed sediment lying on the bluff face, once it reaches some critical thickness and water content, 2) the slip of thawed, thin sheets of sediment along the frozen/thawed ground interface, or 3) the slumping of frozen blocks on slip surfaces lying within thaw zones in the otherwise frozen mass (e.g. Eardley 1938, Sigafoos and Hopkins 1952, Mackay 1966, McRoberts and Morgenstern 1973, 1974a,b, Lewis and Forbes 1974, Miles 1977, Harper 1978a, Tomirdnaro and Ryabchun 1974, Church and Miles 1983).

Figure 70. Partly draped vegetation mat and intact blocks of sediment with intact vegetation cover which have toppled or slid from eroding perennially frozen bank along a northern river. Vegetated blocks partly submerged in water protect bank from currents, but exposed bank sediments continue to thaw and degrade under warm air temperatures and solar radiation effects.

Flow processes, such as debris or mud flows, skin flows or slurries are also commonly observed on thawing bluffs, particularly those of shallow angle or where the bluff's base is of a sufficiently low angle to allow accumulation of thawed material as it is sloughed from the upper, steeper face (e.g. Eardley 1938, Tomirdnaro and Ryabchun 1974, Are 1977, McRoberts 1978). Gravitational flows are particularly important in removing thawed bluff sediments and thus maintaining the angle of the slope and the exposure of the frozen ground to air and sun (e.g. McDonald and Lewis 1973).

Failed material that is not removed protects the upper beach and lower bluff materials from both thermal and mechanical erosional processes. A gradual lowering of the slope angle results as the upper bluff continues to thermally and mechanically degrade (e.g. Ritchie and Walker 1974, Miles 1977). Walker (1983) found that a single cycle of undercutting, collapse, protection and some upper slope angle lowering was repeated once each season on the Colville Delta.

At locations where ice wedges occur in bluff sediments, they preferentially melt faster than the surrounding frozen sediments and lead to gullying (Fig. 70) (e.g. Lewis and Forbes 1974). This gullying disrupts the thermal regime of adjacent sediments which may then undergo thaw subsidence. Further erosion of these inland bluff sediments can result from the slump and flow of gully sidewall material, as well as the deepening of the

gullies by meltwater flow. Water may seep directly from the ground as it thaws, thereby decreasing its stability and providing a continuous flow within the gully channel (Tomirdnaro and Ryabchun 1974).

Erosion of beach and bluff sediments may therefore disrupt the thermal regime of frozen ground landward of the immediate reservoir environment, and induce accelerated physical degradation and the development of thermokarst in land outside the reservoir's shore zone. This accelerated degradation and expansion of thermokarst is clearly not directly related to mechanical erosional processes in the shore zone but result indirectly from it (Lawson 1983a). Parameters affecting the thermal and physical stability of the permafrost are important in determining the extent of this accelerated development of thermokarst.

MATERIAL ORIGINS

It has long been emphasized by Terzaghi and his students that site geology, including rather minor geologic features, can often be the most important factor governing the performance of engineering structures as well as slope stability (e.g. Terzaghi 1929, 1955, Terzaghi and Peck 1967). Knowledge of geological processes for interpreting site geology is important for understanding spatial variability in geotechnical properties and thus potential engineering problems. Terzaghi's thoughts and experiences apply equally well in analyzing the response of developing shore zones to the littoral processes of reservoirs. Although I have discussed some aspects of erosion in relation to material types and structural features, the importance of their geologic origin needs further discussion.

Alluvial stratigraphic sequences

Surficial materials surrounding northern impoundments are mainly either alluvial or glacial. Alluvial materials were generally deposited by the river that was dammed to form the impoundment or its predecessors. In the older river systems, which were draining the glaciers and continental ice sheets during the Pleistocene age, large quantities of sand- to gravel-size particles were deposited and aggraded on the active floodplain. These deposits were subsequently eroded as glaciers waned, and the sediment and water output decreased. Similarly, rivers were typically braided during high discharge periods, but gradually most of them changed to the meandering systems of the present. Depositional sequences in alluvial valleys reflect these changes, as is well illustrated by the Mississippi River system (e.g. Fisk 1952, Turnbull et al. 1966, Schumm 1971).

These alluvial deposits affect modern-day river activity and morphology, including which reaches are characterized by long-term instability and lateral bank erosion and migration (e.g. Fisk 1947, 1952, Krinitzsky 1965, Schumm 1971). Historical changes in land use and construction practices can cause changes in hydraulic and geomorphic factors that in turn result in channel erosion and adjustment (e.g. Whitten and Patrick 1981). Brice (1964), in fact, concluded that bank erodibility is the most important single variable affecting channel pattern, with erodibility mainly related to the size distribution of bed and bank material (Schumm 1960), although the type of vegetation (Mackin 1956, Smith 1976) and the presence or absence of significant cohesion are also important variables.

Within reservoirs, the response of alluvial deposits to the tractive forces exerted by waves and currents can be expected to be similar. Kachugin (1966), for example, concluded that rates of erosion of different materials could be related directly to wave energy and shoreline morphology. Alluvial and eolian deposits were considered most easily eroded, while certain glacial deposits, particularly clay-rich tills, were least erodible. Stanley et al. (1966) and van Everdingen (1969) additionally emphasized the importance of fractures, joints and bedding planes that act as planes of weakness and decrease the stability of shore zones.

The depositional sequences and their spatial variability in braided and meandering rivers have been described in some detail (e.g. Sündborg 1956, Leopold and Wolman 1957, Williams and Rust 1969, McGowen and Garner 1970, Jackson 1976, 1978, Cant and Walker 1978, Miall 1977, 1978, Rust 1978, Reineck and Singh 1980). Certain repetitive features in these sequences can be related to shore erosion and stability.

In general, the depositional subenvironments composing an active river system floodplain develop characteristic vertical sequences with a predictable spatial distribution. For example, in the classic meandering river, coarse gravels are deposited in the river thalweg, with coarse sand composing the remainder of the bed. Point bars that develop on the inside bend of meanders overlie channel deposits and consist of coarse sand near the base that gradually fines toward the point bar surface. Fine-grained deposits (silt or silty clay) may be deposited within abandoned channels and swales but are mostly deposited in levees and flood basins during flood stage. The result is a nearly consistent vertical sequence with coarse cohesionless sediments at the base and gradually finer cohesionless sands toward the top. Cohesive materials cap the sequence. Several such sequences may compose thick alluvial valley fills, some of which may have been truncated by fluvial erosion.

Although relatively simplistic in approach, this sequence of alluvial deposits is often assumed, when analyzing bluff or bank erosion, to consist of mainly two units: 1) an upper cohesive fine-grained unit and 2) a basal unit of interstratified sands and gravels (e.g. Turnbull et al. 1966, Brunsden and Kesel 1973, Scott 1981, Thorne and Tovey 1981).

Certain repetitive types of erosional processes and failure modes are recognized for this simplified sequence. Typically, entrainment of the lower cohesionless material by currents and waves is more rapid than that of the upper cohesive material, causing it to become undermined and oversteepened. The lower sandy bluff materials erode gradually and continually by sloughing. The upper cohesive bluff material fails by a cantilever block failure as described in a previous section. The failed block then protects the lower part of the bluff from erosional processes until it is abraded and transported away by waves and currents (e.g. Fisk 1952, Thorne and Tovey 1981). Slip failures of the cohesive upper unit result from progressive oversteepening by current erosion; discontinuities in the cohesive sediments appear to act as planes of weakness for such failures (Stanley et al. 1966).

Brunsden and Kesel (1973) similarly found that erosion and recession of bluffs composed of alluvial deposits varied with lithology. The cohesive cap exhibits more resistance, while free face cohesionless material

failed more rapidly by processes such as soil falls, block glides and slumps, the latter being initiated by toe erosion or ground water seepage. For bluffs with a complex interstratification of cohesive and non-cohesive units, cohesive strata act as resistant "hard" points, forming benches and steep vertical faces in the bluffs between cohesionless sand layers and, when at the shoreline, protruding into the water in a scalloped form (Fisk 1952, Brunsden and Kesel 1973). Subaqueous failures caused by scour in the thalweg also initiate sloughing of overlying bank and bluff sands (Fisk 1952, Turnbull et al. 1966). In areas of submerged thin layers of cohesive material, these layers failed along numerous small slip surfaces but for thicker layers, flow, partial liquefaction or shear type failures took place (Turnbull et al. 1966).

Glacial stratigraphic sequences

Glacial deposits are common constituents of northern reservoir shores yet have received scant attention in studies of erodibility. Depositional sequences of single or multiple glaciations are complex and can often consist of multiple, interbedded layers with a highly variable spatial distribution. Such sequences are also often characterized by discontinuous lenses and layers which range in thickness from a few millimeters to well over several meters. Sediments composing them may range in size and gradation from sorted silt- to large gravel-size material, to an unsorted, mixed clay-to gravel-size and structureless sediment that is typically called till. These localized variations in stratigraphy and structural discontinuities, which are also common, can act as controls on the rate, location and style of erosion (Fig. 71).

Consistent repetitive sequences are not common in glacial deposits. Geotechnical properties, such as grain size, moisture content, density, degree of consolidation, permeability, shear strength and Atterberg limits, are thus quite often highly variable within individual outcrops as well as within a region (e.g. Elson 1961, Norris 1962, Kazi and Knill 1969, Grisak et al. 1976, McGown et al. 1978, Fookes et al. 1978, May and Thomson 1978, Thomson et al. 1982, Lutenegger et al. 1983). In general, such properties are related to the origin of the glacial deposits (e.g. McGown 1971, McGown and Derbyshire 1977, Boulton and Paul 1976, Edil et al. 1977, Lutenegger et al. 1983). Unfortunately, although the complexity of the geotechnical and hydrologic properties of glacial sediments has been recognized for some time (e.g. Terzaghi 1955, Elson 1961), the capability does not exist for predicting this variability. Even less information is available on predicting the response of glacial materials to erosional forces (e.g. Quigley and Zeman 1980).

Studies of sedimentation by active glaciers over the last 15 years or so have significantly increased knowledge of the multiple origins of till and other glacial deposits, showing that the traditional view of the two origins for till (lodgement and ablation) (e.g. Flint 1971) is overgeneralized (e.g. Boulton 1972, 1976, Lawson 1977, 1979, 1982b, Shaw 1977a,b, 1979, 1980, Eyles 1979, Robinson 1979). The multiple origins for till and other glacial sediments are important in assessing its sedimentologic variability, since its genesis and post-depositional history clearly affect the geotechnical and hydrologic properties of glacial sequences.

Figure 71. Eroding slope composed of complex glacial deposits in Iowa with numerous sand lenses interspersed in diamictons that acted as groundwater sources for springs. Piping, subsurface erosion and collapse and flowage of eroded material ensued.

Sediments may be deposited in a variety of locations in the glacial environment, including subglacially beneath the glacier's sole, supraglacially along and on top of the active or stagnant ice margin, and proglacially at some distance from the active ice margin. A variety of distinct depositional and erosional processes are active within each region. The proglacial region is typically dominated by erosion or deposition by braided outwash streams. These deposits, while coarser overall, are similar in their geotechnical properties to the alluvial deposits discussed previously and will not be considered further here. Proglacial fluvial processes and deposits are discussed by Hjulström (1952), Krigström (1962), Rust (1972, 1978), Church (1972) and Boothroyd and Ashley (1975).

Sedimentation processes within the subglacial and supraglacial environments are basically of two types -- 1) primary processes that release and directly deposit the debris in transport within the glacier ice, and 2) secondary processes that rework and redeposit this debris or glacial sediment. Primary processes may include subglacial deposition of debris directly from active, debris-rich ice, usually called the lodgement process, or the in-situ melting of stagnant, debris-rich ice or the melt-out process. Resedimentation commonly results from sheet and rill flow of meltwater, spalling or slumping of sediments on buried glacier ice, gravitational flow of sediments overlying or adjacent to glacier ice, falling, sliding or rolling of sediment as it ablates from active or stagnant ice, settling of debris through water after melting from ice overlying subglacial lakes, and deposition of suspended sediment from proglacial and supraglacial lake waters.

Physical properties of deposits from primary processes mainly result from the mechanics of the glacier and the debris source, whereas those of secondary processes are newly developed by those resedimentation processes (Lawson 1979, 1981b). Post-depositional processes after glaciation has ended may modify certain properties, but otherwise the principal attributes generally remain distinct (e.g. Quigley 1975, Eyles and Sladen 1981).

Primary and secondary processes can each produce till-like sediments that are characterized by poorly sorted admixtures of clay- to boulder-size particles. These till-like materials are referred to descriptively as diamictons (Flint 1960). The term diamicton is used as the equivalent of descriptive grain size terms such as silt, sand and gravel, and is preferred over the genetic term till, which implies an origin, if detailed studies have not been undertaken to define the actual process of deposition at that particular location (e.g. Lawson 1979, 1981a, Eyles et al. 1983). Detailed study of an assemblage of sedimentologic properties and stratigraphic relationships is usually required to define diamicton origins (e.g. Boulton 1976, Lawson 1977, 1979, 1981a, Shaw 1977a, 1980, 1982). Such studies require a careful analysis and are quite often time consuming. Because of the similarity of diamictons of different origins (they may be deposited, for example, by mud or debris flows, turbidity currents or colluvial processes, as well as in several different depositional environments), precise interpretation of their genesis is not always straightforward (e.g. Dreimanis 1976, Haldorsen and Shaw 1982).

The variability in geotechnical properties of glacial diamictons makes it important to use descriptive terms in initially assessing the response of glacial deposits to erosional processes. Where the processes of deposition can be identified, genetic terms can be applied and may then be useful in identifying specific geotechnical and sedimentologic properties common to diamictons of that origin. Research results to date are, however, insufficient to have confidence in making such correlations.

Nevertheless, attempts have been made to develop depositional models that relate processes and landforms, and hence variations in geotechnical properties, on a regional scale (Boulton and Paul 1976, Boulton 1978, Eyles and Sladen 1981). The use of such proposed models for engineering purposes in North America has thus far been strictly limited. In the case of shore erosion, it may be more profitable to consider detailed representative vertical and lateral sequences of deposits for typical glacial depositional environments. Unfortunately in this regard, existing data are insufficient and each site must be examined separately in detail.

Kemmis et al. (1981) found that depositional sequences in Iowa could be separated genetically into two units that related well to gross changes or differences in their sedimentologic and hence geotechnical properties, including shear strength, hydraulic conductivity, and compressibility. This bipartite arrangement provides a means of readily identifying diamicton units of different origin for engineering or agricultural purposes (Lutenegger et al. 1983).

Kemmis et al. (1981) concluded that the two units had broadly different origins and therefore could readily be distinguished. The lower unit consists of a rather uniform, unstratified basal diamicton that was deposited mainly by subglacial processes, while the upper unit consists of high-

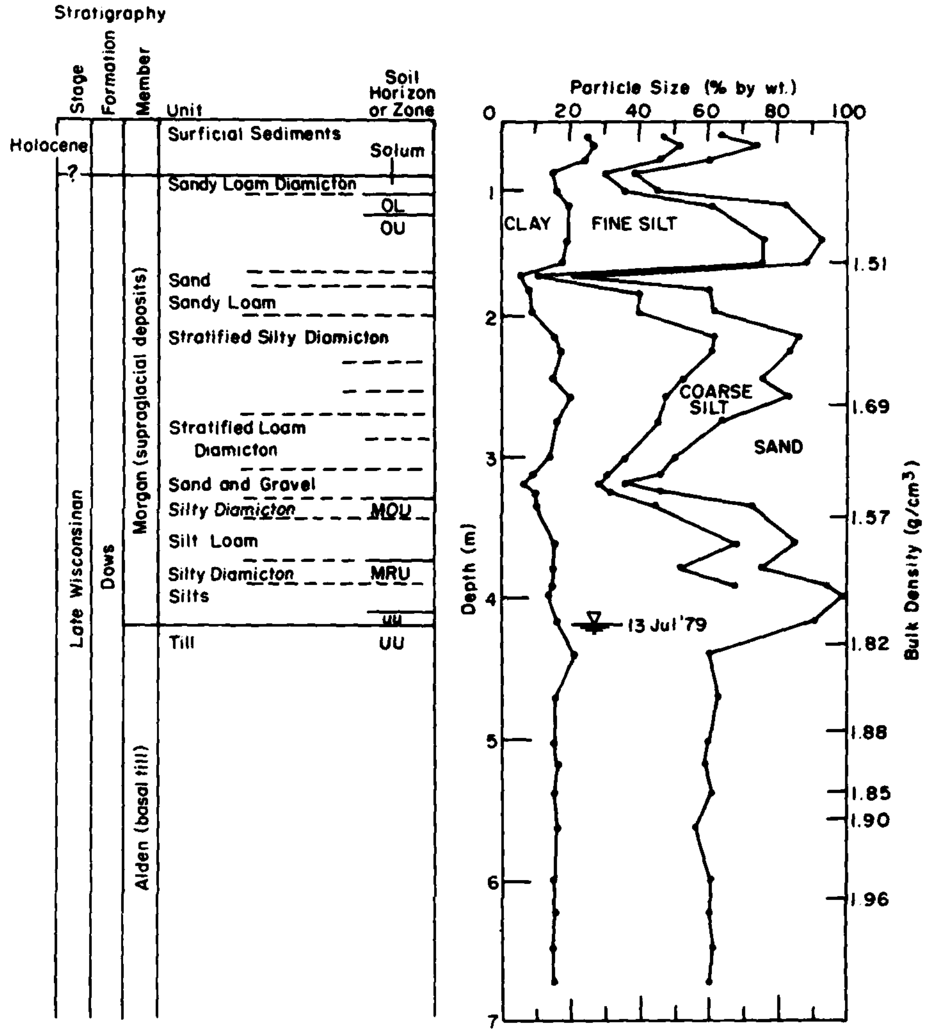


Figure 72. Stratigraphy and textural data for borehole core consisting of the supraglacial and basal units of Kemmis et al. (1981).

ly variable sediments deposited by supraglacial processes. This upper unit was deposited by a variety of resedimentation processes including sediment flows, slumps, and meltwater flows and other processes acting on or adjacent to glacier ice. These processes interact complexly with one another to produce an interbedded, stratified sequence of sorted and unsorted materials that include multiple diamicton layers. The basal diamicton is commonly interpreted as a basal till.

The stratigraphy and trends in grain size typical of these units are shown in Figure 72. Basal and supraglacial units are differentiated on the basis of matrix texture, nature of stratification and sorting, stratigraphic position and nature of contacts between units, and geotechnical properties including density, consolidation data and consistency (Kemmis et al. 1981). Additional data on sedimentary structures in exposures are also incorporated where possible.

Figure 73. Basal unit of Kemmis et al. (1981) showing generally structureless and uniform appearance, with clasts dispersed randomly within a finer-grained matrix. Large gravel clasts (photo center) mark base of deposit.

For the Iowan sequences, the basal unit is typically uniform in texture, lacks interbedded sorted deposits, is firm to very firm in consistency, exhibits considerably higher density (1.8-1.97 g/cm^3) and generally tends to be moderately to highly overconsolidated (Fig. 73). Supraglacial deposits, in contrast, are characterized by textural heterogeneity, with interbedded and often discontinuous layers and lenses of diamictons, silts, sands and clays. The supraglacial unit as a whole is of variable consistency, but generally friable to loose, of moderate bulk density (1.5-1.7 g/cm^3), and variable in the degree of consolidation, but typically normally consolidated (Fig. 74). In addition, sedimentary structures are often observed in the sorted silty and sandy deposits of the supraglacial unit.

The distinctions described by Kemmis et al. (1981) and Lutenegger et al. (1983) are important for erosion studies because they recognize that geotechnical properties differ with the origin of the materials. Thus, they provide an initial means of separating distinctly different deposits that result from two unique styles of deposition in different parts of the glacial environment, without actually requiring the identification of the specific origin of individual layers or beds. The susceptibility of depositional sequences composed of supraglacial sediments to tractive erosion

Figure 74. Typical heterogeneous supraglacial deposits in Iowa consisting of interbedded diamictons, silts and sands, with numerous lenses and discontinuous layers of well-sorted sands and gravels.

and instability is generally higher than that of the basal unit. This general difference may provide an initial means of assessing shore erosion in glacial terrain. Further detailed field studies of glacial sedimentation are needed to progress beyond this level of interpretation on a regional or local basis.

Lateral and within-site variability in recession rates, and thus apparently resistance to erosional processes of shore zones composed of glacial materials, has been observed by Christopher (1959), Gelinas and Quigley (1973), Lineback (1974), Edil et al. 1977, Calkins et al. (1978), Gray and Wilkinson (1979), McGreal (1979), Black (1980) and Birkemeier (1981), among others, with certain distinctive effects of stratigraphy, weathering and anisotropic features in individual beds of glacial sequences on erosion described by Quigley (1975), Sterrett (1980) and Sterrett and Edil (1982). In contrast, Buckler and Winter (1975) found no relationship between bluffs partially composed of glacial deposits and retreat rates. As stated previously, stratigraphy clearly may affect ground water conditions and thus the occurrence of erosion by seeps, piping and related overland flow, and the stability of bluff and beach zone materials (e.g. Grisak and Cherry 1975, Sterrett and Edil 1982). Jointing and other features unique to certain of the glacial diamictons may likewise have overriding effects on stability in the shore zone, yet are relatively unstudied (e.g. Kazi and Knill 1973, McGown et al. 1974, Eyles and Sladen 1981). Further studies are needed of eroding shores composed of glacial deposits.

ICE COVER

An ice cover has mainly two effects on the shore zone. It may act to protect the beach and lower bluff zone sediments from attack by waves, currents and subaerial processes, or it may actually cause erosion as the result of its impingement upon beach and bluff zone sediments. Literature on certain aspects of ice cover erosion or protection of inland shore zones was recently reviewed by Dionne (1979) and Gatto (1982b, 1984), while Michel (1970) has reviewed the engineering aspects of ice cover motions, and Kovacs and Sodhi (1980) and Sodhi and Kovacs (1984) have reviewed the occurrence and mechanics of ice pile-up and ride-up along coasts. These references should be consulted for further details.

Erosion results from 1) winds that drive the ice cover up, across or along beach and bluff sediments, sometimes overtopping bluffs as high as 8 to 10 m and driving it inland up to 100 m (Sodhi and Kovacs 1984), 2) ice expansion against shore zone materials as the ice cover forms and grows, and 3) a vertical motion of the ice cover on the shore face as the water level rises or falls. Ice expansion appears important mainly in smaller lakes and impoundments, and in bays of larger water bodies (e.g. Pessl 1969, Worsley 1975, Dionne 1979). Wind-generated shove of ice requires a sufficiently long fetch for wind stresses to be large enough to force ice across beaches (Fig. 75). The ice cover must also be incomplete, with open water around it. Even after freeze-up, a rise in water level can detach and float the ice cover within open water, making it susceptible to the winds of winter storms. Storm-generated waves or surges in water level

Figure 75. An ice cover on Orwell Lake that was shoved onto beach sediments and then broken up as water level in the reservoir dropped.

(seiching) may drive fragments or complete ice covers inland, well beyond the shore zone (e.g. Croasdale et al. 1978, Kovacs and Sodhi 1980). Similarly as temperatures rise in spring, melting commonly occurs more rapidly along the shore than within the lake (Williams 1966), thereby detaching the ice cover and permitting winds and currents to move it about. The destructive action of ice shove in spring may be limited, however, even though ice is thickest then, because its strength can be severely reduced by candling, which results from selective melting along ice grain boundaries after air temperatures rise above freezing (Gow and Langston 1975).

The occurrence of ice cover expansion onto shore zone materials has been recognized for some time (e.g. Buckley 1900, Laskar and Strenzke 1941, Rose 1946, Zumberge and Wilson 1952, 1953, 1954, Nichols 1953, Montagne 1963, Wagner 1970). Various factors identified as affecting the magnitude of ice expansion include air temperature, the solar energy absorption and temperature of the ice cover, the thermal gradient, thermal expansion and rheology of the ice, the presence of water-filled cracks, and shoreline morphology (Michel 1970). Ice ramparts are produced by infilling of tension cracks, which form during a rapid fall in air temperature, with water that quickly freezes to increase the mass of the ice cover. A subsequent rise in temperature results in expansion of the ice cover, which, because its surface area is greater than before, exerts a compressive force toward and against the shore zone (Michel 1970).

A number of researchers have calculated the pressures that may be exerted by an expanding ice sheet, mainly in relation to its potential effects on dams or other structures (e.g. Michel 1970, Korzhavin 1971). For example, Korzhavin (1971) defined a widely used equation for calculating the crushing pressure against a structure that is based upon the compressive strength of the ice, the amount of ice in contact with the structure and the structure's shape.

The abrasive action of ice during shove or movement against beaches and bluffs is limited by the maximum local pressure the ice can withstand before it fails. This limit is generally thought to be the "crushing strength" or force, if the ice cover is relatively thick (e.g. Michel 1970), or its buckling force if the ice cover is relatively thin (e.g., Assur 1971, Croasdale and Marcellus 1978, Kry 1980, Kerr 1981). Typically for thin ice, a bending failure against the shore face quickly results, and fragments of the ice cover can build up in rubble piles (e.g. Croasdale and Marcellus 1978), particularly during early freeze-up when the ice cover is incomplete and repetitive onshore movements take place (Cox et al. 1983).

In the situation where the ice cover is incomplete, but frozen to shore sediments, Croasdale and Marcellus (1978) concluded that only a crushing failure or ductile flow of the ice will take place since the required forces for such failures are less than those for a bending and buckling failure, or failure and slip along the forzen bond between the ice and sediments. Complex situations involving each mode of failure have been described from various coastal sites (Gladwell 1977, Kovacs et al. 1982).

The effects of a fragmented ice cover that is pushed along or against the shoreline are more difficult to predict because of the inherent variabilities in physical properties of the ice. Thus the local distribution

of normal and shear stresses at the shoreline (e.g. Stewart and Daly 1984, Sodhi and Kovacs (1984)are also quite unpredictable.

Equations for calculating the stresses necessary to cause buckling, crushing, ice/sediment bond, and ductile failures at the shoreline are presented by several authors, including Croasdale and Marcellus (1978) and Sodhi and Kovacs (1984). Theoretical expressions for estimating forces generated during ice pile-up or ride-up suggest they may range from about 10 to 350 kPa ($\approx$ 1.5 to 50 psi) (Kovacs and Sodhi 1980). However, it remains impossible at present to predict when and where such forces necessary to cause ice ride-up or pile-up will occur.

The onshore movement of ice by the wind and its pile-up and jamming on the beach or against bluffs are controlled by a number of factors. Normally as an ice cover moves across the ground surface, it reaches a point of instability and breaks into pieces that can, in turn, be shoved ahead of the intact ice sheet or they can pile-up in mounds or ridges approximately paralleling the shoreline.

The force exerted by the ice push is a function of the 1) ice cover thickness and size, 2) magnitude of the wind stress, 3) effects of nearshore currents and waves, and 4) ice strength and modulus (Croasdale et al. 1978, Sodhi and Kovacs 1984). Ice strength in turn depends upon its thermal and structural state, the latter being affected by crystal structure, strain rate, flaws, such as cracks and candling, or the inclusion and volume of air trapped within it (Gow and Langston 1975, 1977).

The resistance offered by the shore to the ice force is determined by 1) slope angle(s), 2) slope height, 3) slope morphology, 4) offshore bathymetry and nearshore profile, and 5) frictional resistance of the ground surface (e.g. Croasdale et al. 1978, Sodhi and Kovacs 1984). In addition whether the ice is initially frozen to the bed or floating free, and the level of water relative to beach and bluff configuration will vary the resistance (Sodhi and Kovacs 1984).

The distance of ice ride-up is therefore limited, for example, by low ice strength, steep slopes and high bluffs, a rough ground surface, and thin ice of limited extent. Theoretical calculations and observations also indicate that jamming and a pile-up of ice fragments can be initiated by a rapid change in slope angle (e.g. Kovacs and Sodhi 1980). This jamming would actually limit the potential effects of ice shove. In addition, along-shore variations in onshore movement, failure mechanisms and thus ride-up or pile-up result from the inherent natural variations in the motive and resistive factors discussed above, as well as shoreline configuration. Kovacs and Sodhi (1980) present various theoretical expressions for relating the resistive and motive factors for several different conditions of ice ride-up and pile-up.

The actual erosion of material or shoreline recession caused by ice covers has not been quantified (Gatto 1982b). Virtually no measurements have been made, although observations clearly indicate that large ridges can be pushed ahead of the ice face, and normally stable beach or bluff sediments can be gouged by the impinging ice cover (e.g. Zumberge and Wilson 1952, 1953, Seibel et al. 1976, Mackay and Mackay 1977, Sommerville and

a. Shorefast ice containing coarse sand and gravel.

b. Broken ice cover fragments showing layering of debris-laden ice containing sediments up to gravel-size.

Figure 76. An ice cover lying on Orwell Lake beach sediments.

Burns 1968, Pessl 1969, Adams 1977, Dionne 1979, Kovacs and Sodhi 1980, Pyokari 1981). Large boulders and aggregates or clumps of beach material including vegetation are readily moved during ice shove and ride-up (e.g. Barnes 1982, Kovacs 1983). Sufficient removal of material or sufficiently deep gouging of backshore sediments may undermine bluff materials and decrease their stability. In addition, ice thrusting can entrain and transport nearshore bottom sediments landward, onto and even beyond the beach (e.g. Kovacs and Sodhi 1980, Kovacs 1983). Multiple abrasive and ice shove events, as may occur in larger reservoirs that initially develop a partial ice cover, would cause significant erosion due to their cumulative effects.

Further, the freeze-on of the ice cover to the beach or bluff face sediments can entrain and transport these materials offshore when the ice cover pulls away during breakup in the spring (Fig. 76). Deposition of sediment on the ice cover by overland flow processes moving off exposed bluffs or beaches may likewise provide eroded material that will be moved offshore by ice rafting. In addition, ice shove can destroy the vegetation cover and leave shore zone soil exposed to subaerial erosional processes, particularly rainfall, runoff and repetitive freezing and thawing.

Gatto (1982b) concluded from his literature review that the effectiveness of an ice cover in causing erosion in reservoirs is directly related to water level. If it is high enough, ice can impinge directly on the bluff face; if it is below the normal water line, it may have no direct effect at all. Lowering the water level in a reservoir once the ice cover has formed, will therefore minimize any effects of ice expansion or shove (Rose 1946). Rapid transitory vertical movements in the ice cover, such as observed by Wuebben (1983a,b) during ship passage, may similarly result in impingement of ice against beach or bluff face materials, depending upon the water level before passage. A frozen condition could minimize the effects of ice ride-up or ice shove, since this condition generally decreases erodibility. Additionally the degree of ice cover attachment and its strength, rate and range of fluctuations in water level, general mobility of the ice cover, and nearshore water depth will affect the intensity of erosion.

A complete ice cover along the shore protects the beach and bluff from the direct attack of waves and currents, and can dampen waves even when incomplete if the ice is situated along the shoreline (e.g. Brochu 1961, Ouellet and Baird 1978, Coakley and Rust 1968, Avakyn 1975, Abele 1977). Similarly the buildup of ice that either forms on or is shoved onto shore early in winter could remain in place and protect beaches late into spring, especially in northernmost regions (e.g. Evenson and Cohn 1979, Kovacs et al. 1982). Birkemeier (1981) measured overall rates of erosion and recession on Lake Michigan and concluded that erosion was most active just before and after nearshore ice formation, when winter storms were active but the associated wave activity was not modified or quelled by an ice cover.

A large reservoir may also develop an ice foot by repetitive wave spray freezing to upper beach and bluff sediments, a phenomenon especially common in large lakes following winter storms (e.g. Zumberge and Wilson 1954, O'Hara and Ayers 1972, Evenson and Cohn 1979). This buildup of ice would protect beach and bluff sediments. Gatto (1982b) concluded that an ice foot is uncommon in many small reservoirs because of the relatively rapid growth of the ice cover early in winter.

The extent and nature of sub-ice processes (including erosion by currents and waves in the lacustrine environment) is virtually unknown, but intense physical reworking and transport of nearshore materials of the lake bottom are possible (Rea et al. 1981). Wuebben et al. (1978) found that the passage of large vessels in connecting channels between the Great Lakes caused significant erosion and transport of bed material beneath an ice cover. Ship passage results in a drop in the wat r level and ice level; this translatory movement of the water and ice generates currents that can entrain and transport the bed material as bed load and suspended load. In addition, rapid changes in stress at the bed can also induce a spontaneous, explosive liquefaction which thrusts sediments into the water column. This situation may be analogous to that where intense wind waves generated in open offshore areas of an impoundment move into and beneath ice-covered, nearshore areas.

MODELS OF SHORE ZONE DEVELOPMENT AND EROSION

The attack of beach and bluff sediments has usually been quantitatively related to the dissipation of energy from waves and related currents in the nearshore zone (e.g. Rossmann and Seibel 1977, Miller 1976, Kondratjev 1966, Kachugin 1966, Quigley and Gélinas 1976, Sunamura 1977, 1982a and b, Black, 1980, Carter et al. 1981), although the relationship of wave energy to the rate and amount of bluff erosion has not yet been satisfactorily established (Edil and Vallejo 1980). No analyses have quantitatively defined the importance of wind waves as an erosional process in relation to other processes that cause shoreline recession at a specific site, or along a reach of shore, within artificial impoundments or lakes.

Under the assumption that wind waves cause most, if not all, significant changes in the shores of Siberian reservoirs, Kondratjev (1966) proposed a "stable shelf" concept for analyzing bank formation in newly created reservoirs. This concept is similar to that of Bruun (1962) in suggesting that eroded sediment is transported offshore but deposited sufficiently close to the shore to form a protective shoal (Fig. 77). Wave-generated currents caused by strong onshore winds transport this sediment as suspended and bed loads after its mobilization by waves. Erosion ceases after the shelf reaches a width B over which all available wave energy will be dissipated before reaching the backshore and bluff zone.

The morphology of the stable shelf is defined empirically by Kondratjev (1966) as

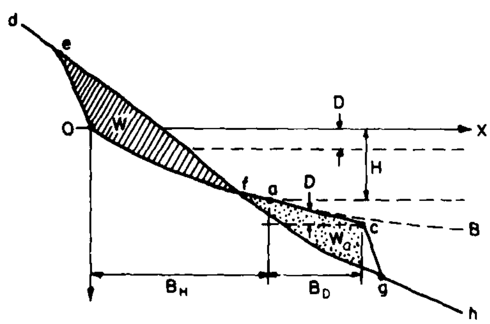


Figure 77. Kondratjev's (1966) conceptual stable shelf model. Parameters defined in text.

$$X = AY^2 + \frac{1}{m_n} Y \tag{70}$$

$$\text{where } A = \frac{m_n - m_o}{20m_n \cdot m_o}, \tag{71}$$

and m_n is related to the slope of the shelf near the shoreline, and m_o is related to the slope at the outer edge of the shelf where wave action ceases. Values for m_n and m_o are given by Kondratjev. Y is shelf height between the water's edge and depth H, at which erosion starts, and is dependent upon wavelength, wave amplitude and the nature of the shelf sediments. X is the stable shelf width.

The width of the shelf B_H at some time t is therefore defined by

$$B_H = AH^2 + \frac{1}{m_n} H \tag{72}$$

with parameters shown in Figure 25. If the water level decreases, this modifies B_H by

$$B_D = D(2AH + \frac{1}{m_n}) \tag{73}$$

where B_D is the increase in width of the shelf and D the decrease in water level height on the shore face. Thus, total shelf width would be

$$B = B_H + B_D \quad . \tag{74}$$

Kondratjev (1966) further expands his concept to include wave action, by assuming this is the only erosive force, and its resultant dispersal of energy transports the eroded material. This equation is

$$w = w' \ (1 - 1^{-xt}), \tag{75}$$

relates w, total volume of shore material eroded, to the incremental loss in volume of sediment w', over a given interval of time. x depends upon the resistance of shore sediments to erosion and can be determined by various iterations of the following equation,

$$\Delta t = \frac{\varepsilon B_H w'}{\bar{N}} \ln \frac{B_H - b_n}{B_H - b_{n+1}} \quad . \tag{76}$$

This equation accounts for the volumetric loss of shore material w' over a given time interval Δt and in response to a total wave energy $\bar{N}$ per unit length of shore. b_n and b_{n+1} indicate the width of the developing shelf at the beginning and end of the time interval under consideration. ε represents the resistance of shore materials to erosion by wave action. By varying $\bar{N}$ for various water levels, the total volume of eroded material and the volumetric loss w can be determined. Thus in eq 75, t would equal $\Sigma\Delta t$.

Kachugin (1966) has empirically related the quantity of material eroded Q_e to available wave energy E_w by

$$Q_e = E_w \cdot K_e \cdot K_b \cdot t^b \tag{77}$$

where K_e = wash-out coefficient
K_b = a coefficient of bank height, expressed as $h_b \cdot a$
h_b = average bank height
a = *coefficient dependent* upon bank composition
b = a coefficient related to the formation of shoals from the eroded sediment, which varies with the dimensions and dispersal of eroded sediment in the foreshore.

E_w is determined by wave climate studies and two simple relationships

$$E_i = e_i \sin(\alpha) \tag{78}$$

where e_i is the wave energy for winds of a given bearing α with respect to the shoreline, and

$$E_w = E_1 + E_2 + E_3 \ldots + E_i \, . \tag{79}$$

The parameter a also varies with K_e; values are given by Kachugin (1966). Variations in bank height with time are used to account for water level fluctuations. Thus Q_e is the volume of eroded sediment per unit length of the shore under the total energy of waves attacking the shore zone.

K_e is essentially a measure of the erodibility of bank materials under the action of waves. Readily eroded sediment such as sand and sandy loam have larger washout coefficients than much less erodible cohesive sediment such as clay and clayey silt. Kachugin (1966) presents empirical estimates of K_e for different sediment types.

As with Kondratjev's (1966) concept, Kachugin's model assumes that eroded sediment is moved offshore a distance determined by the strength of the unidirectional nearshore currents and thus wave energy. Longshore currents are equally important in moving washout material away from the eroding shore. Offshore bars develop a distance and depth from the shoreline determined by these factors (Kachugin 1966). Coarser sediment is assumed to be deposited in nearshore shoals that are continually reworked by waves.

Once Q_e is calculated, the quantity of eroded sediment is used to estimate the width of the zones from which shore material is eroded over fixed time periods. Kachugin (1966) used characteristic beach profiles for average bank heights along the shore reaches of interest to estimate graphically the expected shore zone modifications. He indicated good correlation to observed changes in Siberian reservoirs with time.

Van Everdingen (1969) applied both Kondratjev's (1966) stable shelf concept and Kachugin's (1966) wave energy equation and washout coefficient to evaluate erosional modifications to the nearshore zone of Diefenbaker Lake, a valley reservoir in Canada. He defined typical shore profiles for

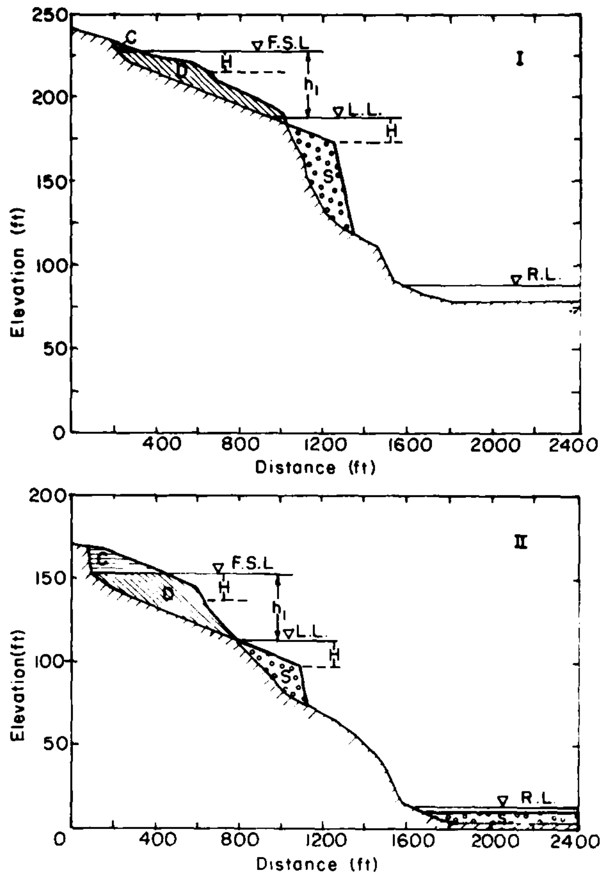


Figure 78. Erosion and sedimentation of the west bank of Diefenbaker Lake for a cross section through a headland cliff (I) and and a gully (II). Calculated extent of erosion above the full water level (F.S.L.) is shown by C and between F.S.L. and the low level (.L.) is shown by D. S is the eroded and redeposited sediment derived from erosion. R.L. is the original level in the river before damming (after Van Everdingen 1969).

the reservoir -- one across a well-defined headland cliff and the other across a gully. Based upon their original configuration, these profiles were then altered to an idealized configuration according to the theoretical models for final full water level (Fig. 78). This permitted theoretical calculation of the amount of erosion and extent of shoreline recession between the low and full water levels, and of erosion and recession above full water level. Anticipated locations of redeposition of the eroded sediment followed that assumed by Kondratjev's model.

The stable shelf over which water level fluctuations will occur (beach zone width) was predicted by Van Everdingen to be up to 215 m wide, with a slope of 3°. Furthermore, most significant changes (90%) to the shore zone were predicted to occur in 5 to 10 years. Calculated changes in storage capacity (volume) as the result of erosion and redeposition of these sediments revealed a decrease in total storage capacity, both at high (-0.25%) and low (-4%) water levels, but an actual increase in usable storage (+7.4%) for power and irrigation purposes.

An idealized concept of equilibrium profile adjustments as envisioned by Bruun (1962) was applied by Hands (1980) to predict shore profile adjustments and shoreline recession resulting from rising water levels in the Great Lakes. This conceptual model could also be used for roughly estimating the response of the shore profile in mature reservoirs whose full pool height is permanently raised or lowered.

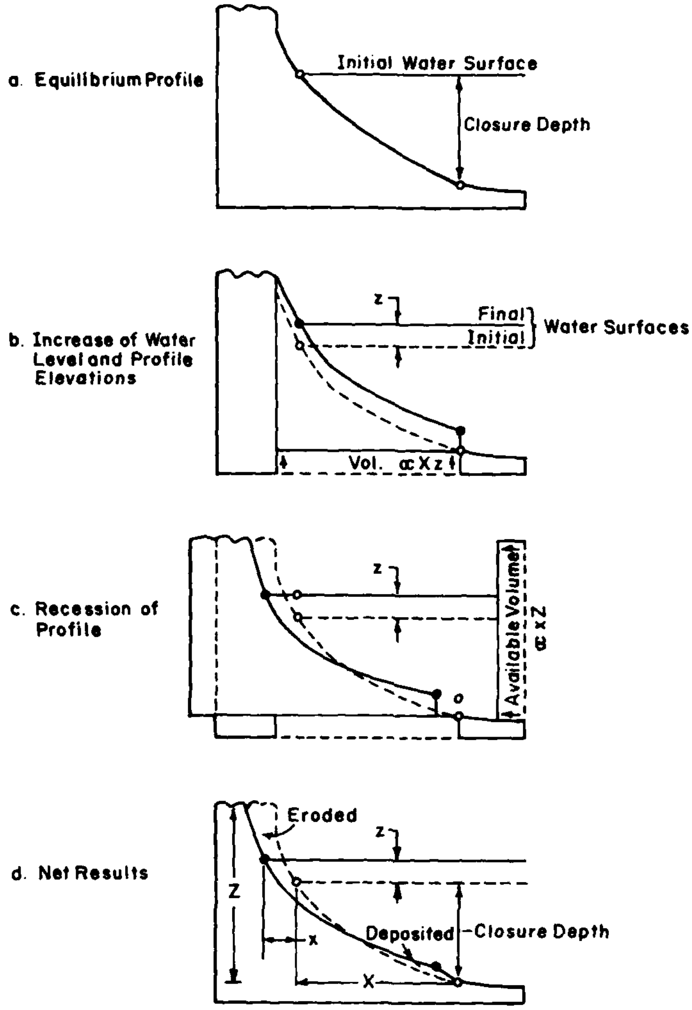


Figure 79. Model of erosion and shoreline recession resulting from a rising water level in the Great Lakes according to Hands (1980). Parameters defined in text.

As water levels rise, erosion of the upper beach and bluff zones results in shoreline recession and adjustment of the profile configuration to a limited depth (closure depth) and distance offshore. The eroded sediment is transported into the immediate offshore zone where it accumulates, resulting in a net rise in the profile (Fig. 79). Ultimately, the overall effect is to reestablish an equilibrium shore profile inland from its original position, at a distance above its original position equal to the net rise in the water level. For falling water levels, the opposite effect occurs and a net accretion in shore materials takes place with an offshore shift in the profile.

In equation form,

$$x = \frac{zX(R_A)^{sg(z)}}{Z} \qquad (80)$$

where X and Z = profile dimensions
z = the change in water level
x = the net lateral retreat of the shore (Fig. 79).

The relation sg(z) = 1 if z > 0 and water level is rising, or sg(z) = -1 if z < 0 and the level is falling.

R_A relates to the stability of shore-eroded material in the immediate offshore zone; that is, how much of the eroded material will remain in this zone, and how much will be lost to offshore or longshore transport and thus not contribute to shore profile development. $R_A \approx 1$ when there is no net loss, $R_A > 1$ when there is a loss and additional shore erosion must contribute sediment to compensate for that loss. If a net loss is defined, R_A is reduced proportionally.

The factor R_A is difficult to evaluate on the basis of current knowledge. It can be estimated by defining the textural characteristics of the sediment along the entire profile length to the closure depth (Fig. 79), and also by evaluating the effects of currents, waves, and other processes that cause sediment movement and redeposition. Storm frequency and intensity also need to be considered. Thus, determining R_A requires detailed field studies and subjective appraisal of all available data.

Hands (1980) discusses in some detail the difficulties in applying this model to the natural environment. Particularly important are difficulties in obtaining accurate bathymetric surveys in order to define the closure depth. Also, defining R_A requires an accurate knowledge of the actual volume lost, whether it be offshore, onshore, or alongshore. As Hands points out, application of this concept must be tempered with sound engineering and geotechnical judgment for the shore reaches under study.

According to Sunamura (1977, 1982a,b), long-term rates of sea cliff erosion are quantitatively related to the erosive force of waves. Based upon laboratory wave tank experiments that were confirmed by previous field and laboratory data, this proportional relationship is expressed as

$$\frac{dx}{dt} \propto F = \ln \left(\frac{f_w}{f_r}\right) \qquad (81)$$

where dx/dt is the erosion rate and F the erosive force of waves. F is defined as proportional to $\ln (f_w/f_r)$, where f_w is the erosive force of waves and f_r is the resisting force of the cliff material.

Resistance of rocks to wave erosion is obviously controlled by their mechanical properties and structural features, such as joints or stratification, which can act as planes of weakness. Weathering coupled with wave action will reduce their effective strength or erosive resistance with time. Sunamura (1977) concluded that erosion of rock cliffs by breaking waves and by waves during runup could be expressed by

$$\frac{dx}{dt} \propto \left(\ln \left(\frac{\rho g H}{S_c}\right) + C\right) \qquad (82)$$

where S_c = compressive strength of cliff face rocks
ρ = density of water
H = wave height at the cliff base
g = gravitational acceleration
C = a non-dimensional constant

Both C and the proportionality factor are defined by field data.

Sunamura (1982a) slightly modified this equation to account for short-term recession of bluffs composed of sedimentary rocks over a given time interval, so that

$$\frac{dx}{dt} = K[C + \ln (\rho g H/S_c)] \qquad (83)$$

where K is a constant with units of length over time. Integration of eq 83 defines the total distance of cliff retreat with time. Thus,

$$x = K\left(C + \ln \frac{\rho g H}{S_c}\right) t \qquad (84)$$

can be used to estimate the distance of bluffline or shoreline recession. Furthermore, the critical wave height to initiate erosion H_{crit} is defined as

$$H_{crit} = S_c\, e^{-C}/\rho g \qquad (85)$$

Because variability in wave height with time at a location can be expressed as a frequency distribution, cliff retreat caused by a given group of waves of height H_i is

$$x_i = K\left(C + \ln \frac{\rho g H_i}{S_c}\right) t_i \qquad (86)$$

where $t_i = \delta_i\ \tau$ and
t_i = duration of waves of height H_i
δ_i = frequency of occurrence of waves with height H_i
τ = length of time interval under consideration.

δ_i is simply the number of occurrences of waves of H_i divided by the total number of occurrences of all waves during the time interval.

Under field conditions, the critical wave height H_i is not immediately known. Therefore the total distance of shoreline/bluffline recession can be initially expressed as a summation of the erosion caused by each wave height impinging on the bluff. If H_j is arbitrarily assumed to represent the critical wave height at a location and the waves of a height less than H_j are assumed not to cause erosion, the length of recession is

$$x = \sum_{i=j}^{n} x_i = \sum_{i=j}^{n} K\left(C + \ln \frac{\rho g H_i}{S_c}\right)\delta_i \tau \qquad (87)$$

where n is the largest observed wave height. By repetitively solving this equation with field data for each H_i, the values of C and K can be computed for

$$K = x/\tau \sum_{i=j}^{n} \left(C + \ln \frac{\rho g H_i}{S_c}\right)\delta_i \tag{88}$$

and

$$C = -\ln \frac{\rho g H_i}{S_c} \tag{89}$$

from eq 87 and 83, respectively.

In practice, data on wave heights at the cliff base are not always available. Offshore wave height, however, can be used to estimate H with the empirical relationship

$$\frac{H}{d} = 0.78 \ , \qquad \text{with} \quad d = h + \eta \tag{90}$$

where d = the water depth at the cliff base
h = water depth at still-water level (SWL)
η = water level rise or wave set-up (Fig. 80).

An empirical relationship gives η as:

$$\eta = -\ (3.85 \tan \beta + 0.015)\ h/(1.63 \tan \beta + 0.048)\ H_b \tag{91}$$

where tan β is nearshore bottom slope and H_b is breaker height (Fig. 80). H_b can be estimated from offshore wave height H_o and wave length L_o by:

$$H_b = 0.563\ H_o/(H_o/L_o)^{1/5} \tag{92}$$

While Sunamura developed these equations for eroding cliffs composed of sedimentary rock, the compressive strength S_c, which represents the mechanical strength of the bluff material, could be replaced with tensile strength or shear strength because these indices are closely related and not independent of one another (Sunamura 1982a). Thus, it seems reasonable that substituting a shear stength S_s for S_c may be representative of erosion of bluffs composed of homogeneous, unconsolidated material. Furthermore, bluffs composed of stratified sediments with highly variable structure and composition could be represented by either a smallest value for the weakest material under wave attack, or as a summational effect by

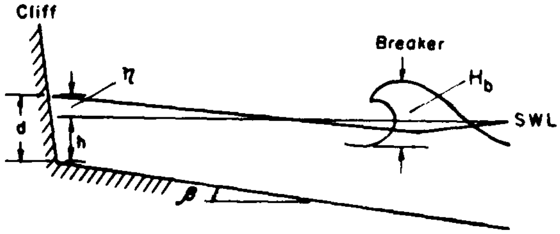


Figure 80. Definition of parameters in Sunamura's (1982) theoretical calculations.

solving eq 87 several times to simulate the variability in response of the individual strata.

Sunamura's (1982a) field observations indicated that eq 83 and 84 adequately represented cliff erosion rates and shoreline recession at two coastal sites in Japan where toe erosion resulted strictly from waves. These bluffs were not subject to large-scale mass movements or other erosional processes to any great degree. At each site, the larger, but less common waves, caused most cliff erosion while smaller waves were important only in removing erosional products that had fallen to the cliff base. His data also suggested that recession rates of beach sediment are apparently related directly to the occurrence frequency of waves above the critical height that caused erosion.

Water level fluctuations and the influence of beach material at the cliff base, both of which may be important in reservoirs are, however, not considered directly by the relationships. In addition, S_c and H did not vary significantly along the shoreline reach under consideration.

Bhowmik (1976, 1978) presented an equation for estimating the significant wave height (H_s) in a reservoir and applied it in analyzing the stability of riprap used for bank protection against wind waves. At a bank location, wave climate is first evaluated with a time series analysis for estimating changes in wave height with time. Bhowmik then evaluated wave climate with statistical techniques, by using an autocorrelation coefficient to define wave periodicity, a Fourier coefficient to define the variance of a series of wave heights from a computed mean height, a spectral density calculation to define wave energy, and a Rayleigh distribution to approximate the distribution of wind-generated wave heights.

Bhowmik derived the equation

$$gH_s/U^2 = 3.23 \times 10^{-3} (gF_e/U^2)^{0.435} \qquad (93)$$

to determine the significant wave height H_s when the wind velocity U, effective fetch length (F_e), and wind direction are known. ($F_e = 1.054\ w^{0.6}\ F^{0.4}$ where w is lake width and F fetch length.) The parameter g is acceleration due to gravity.

Equation 93 is based upon relationships proposed by Sibul (1955), Saville (1954) and Saville et al. (1964) that are described by Bhowmik (1976) and tested with field data collected in the Carlyle Lake impoundment in Illinois. A nomograph was presented for estimating wave height (Fig. 6, in Bhowmik 1978) and close comparison existed between the measured and estimated significant wave heights.

Bhowmik (1978) also analyzed the stablity of single riprap particles against the forces generated by breaking waves (Fig. 81). After analyzing the forces acting on particles, he derived the equation

$$\bar{w} = [kH^3S_s]/[\gamma^2(S_s-1)^3(\cos\alpha - \sin\alpha)^3] \qquad (94)$$

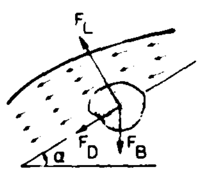
where $\bar{w}$ = the weight of the stable riprap particle
k = a coefficient (values defined in Bhowmik 1978)
S_s = the specific gravity of the riprap particles
γ = the specific weight of water
α = the angle of the beach face
H = wave height.

This expression makes the reasonable assumption that the depth of water at which the wave breaks equals the wave height. Hudson (1959) also made an analysis of wave forces and derived a similar relationship.

Bhowmik (1978, Fig. 8) presents a nomograph based upon eq 94 for estimating the median size particle that is stable under existing wave conditions. Field data from Carlyle Lake, a reservoir in Illinois, was used to verify its validity. He also indicates that an estimate of the width of the bank needing stabilization must be calculated and this estimate depends upon expected low water level and maximum high water level plus freeboard. Freeboard, the highest effective location of wave action (Saville et al. 1962), depends upon wave height, wave runup and wind tide. Wind tide can be estimated by

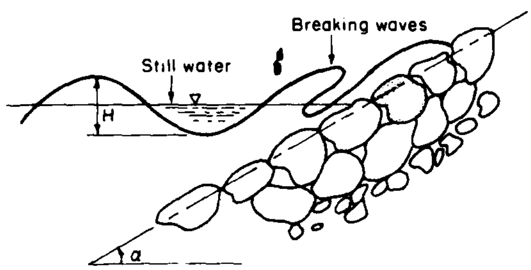
$$S = (KU^2 F_e \cos\Phi)/D \qquad (95)$$

where S = height of wind tide
K = coefficient
U = wind velocity
F_e = effective fetch
D = average water depth
Φ = the angle between the wind direction and the plane of the water surface (Saville et al. 1962).



a. Forces acting on particles with

F_L = lift force
F_B = buoyant weight
F_D = drag force
S_s = specific gravity of stone
$\bar{W}$ = weight of the stone



b. Relationships of wave and riprap particle parameters.

Figure 81. Stability of riprap particles as determined by Bhowmik (1978).

An additional point which Bhowmik (1978) considers is that the effective fetch on reservoirs may be affected to an unknown degree by local topography and vegetation. For example, in low relief areas that lack protection from trees, such as in the Midwest, the effective fetch over which the wind can blow unimpeded may be greater than the simple length and width of the water surface. Conversely, trees surrounding impoundments may reduce wind velocity and locally reduce the water area affected by the wind.

Bhowmik (1978) also quantitatively analyzed waves generated by power boats based upon field measurements of small boat waves in reservoirs. Power boat waves have been cited as possibly important in causing erosion (e.g. Palmer 1973, Williams et al. 1979, Simons and Li 1982). His analysis of these data followed that described above for estimating significant wave height.

Bhowmik's field data indicated that boat-generated waves do not show periodic variations, they are nonstationary in nature, and wave heights are only partly accounted for by a Rayleigh distribution. As might be expected, waves generated by boats close to shore initially produced a peak wave of large amplitude that was followed by small amplitude waves. Waves from boats farther offshore had smaller amplitudes and smoother forms near the shoreline, apparently due to frictional resistance and energy dissipation. In terms of erosion, a much larger amount of wave energy must be dissipated in the beach zone over a shorter time period for waves of boats closer to shore than farther away. Bhowmik (1976) suggested that banning small boats from within 100 ft of the shoreline would minimize erosion caused by boat waves.

The empirical equation based upon the field data for the wave height of boat-generated waves is

$$(H_m/d_s)^2 = (3.45)(10^{-2})(V^{1.174})(X/L)^{-0.915} \qquad (96)$$

where V = boat speed
H_m = maximum wave height
d_s = draft of the boat
X = distance between the boat and wave gauge
L = length of the boat.

Black (1980) examined rates of shoreline recession along Rathbun Lake, a reservoir on the Chariton River in Iowa, by using aerial photographs, bathymetric and beach profile surveys, and maps of the pre-impoundment topography and geology. Assuming a stable slope was already developing below normal pool elevation, he used various standard relationships for wave parameters to estimate future rates of erosion and overall shoreline recession along characteristic profiles that were developed in representative bank materials. Only wind waves were considered an important erosional process. Wind wave parameter relationships and techniques defined by Saville (1954) and the Beach Erosion Board (1962), and estimates of wave run-up following Stoa (1978) were used. Parameters included significant wave height and period, effective fetch and maximum deep water wave height. Black (1980) concluded that the ultimate (final) recession line would involve a 60- to 200-m shoreward movement in the shoreline for the current

pool level management scheme, and require from 20 to 60 years to reach this ultimate shoreline position, the actual length of time depending upon the local shore zone orientation and composition of bluff sediments.

PROCESS INTERACTIONS, ASSESSMENT OF IMPORTANCE, SHORE ZONE EVOLUTION AND RESEARCH NEEDS

It would be particularly useful at this point to present a summary comparing erosion processes, the relative importance of those processes, shoreline and bluffline recession rates, and various factors such as shore zone geology, beach and bluff zone configuration, and thermal and ice conditions for a variety of sites in northern reservoirs. This task is unfortunately not possible given the present qualitative state of knowledge and general paucity of studies of eroding shore zones in reservoirs that are reported in the open literature.

Although it might appear that such a summary should be possible if based upon studies in the lacustrine environment, the large number of variables involved and the limited number of controlled research studies of shore erosion would make this difficult. Data from sea coasts on certain erosional processes are more abundant, but the myriad of possible variables affecting the intensity and activity of those processes, including climate, location (geologically and geographically) and site history, make it difficult to extract useful data for erosional analyses in artificial impoundments.

There are only a very few studies of erosional processes in reservoirs, most being qualitative assessments of which processes are active and what factors appear important (Kondratjev 1966, van Everdingen 1969, Pederson 1971, Coakley and Hamblin (undated), Cyberski 1973, Pulyayevskiy et al. 1978, Adams 1978, Simons et al. 1979, Hagerty et al. 1981, Reid 1982, 1984, Newbury and McCullough 1983). A few authors have attempted to quantify shoreline change in reservoirs through empirical or theoretical calculations, especially when wind-waves are assumed to be the predominant erosional force (e.g. Kachugin 1966, 1970, Bhowmik 1978, Black 1980). Slope stability observations and analyses have also been made by the U.S. Geological Survey along western reservoirs (e.g. Jones et al. 1961, Erskine 1973, Simons and Rorabaugh 1971). In addition, various studies by Corps of Engineers Districts provide qualitative observations on erosional processes and some studies of measured sections to define shoreline recession rates (e.g. St. Paul District 1979, Seattle District 1956 and supplements 1960 to 1974, Omaha District 1965, 1971, 1976, Pittsburgh District 1956, 1960, Seattle District 1971). Recently Gatto and Doe (1983) estimated historical shoreline recession rates along 10 northern reservoirs from aerial photographs and, based upon available data and limited field observations, discussed possible causes of erosion. They strictly qualify their data, however, as to its limited accuracy and lack of on-site, year-round detailed ground truth studies of erosional processes and rates.

The most complete set of studies examined in this review, although still generally lacking in quantitative treatment of the problem, are those of the eroding shores of the Great Lakes, including recent Ph.D. dissertations (e.g. Chieruzzi and Baker 1958, Christopher 1959, Pincus 1962, Seibel

1972, Davis et al. 1973, Maresca 1975, Carter 1976, Berg and Collinson 1976, DuMontelle et al. 1976, Vallejo 1977, Quigley et al. 1977, Schneider et al. 1977a,b, Edil and Vallejo 1977, 1980, Mickelson et al. 1977, Hands 1979, 1980, Sterrett 1980, Birkemeier 1981, 1982).

Still the general conclusion remains that, although the major erosional processes discussed in previous sections and the importance of process interactions and interrelatedness have been identified, few data exist on 1) the relative importance of the process, 2) the quantities of material eroded by a specific process at a particular location and the resulting shoreline change, and 3) how changes in the various parameters affect both the style and intensity of erosion.

Sterrett (1980) and Reid (1984) estimated the percentages of the total amount of material eroded by a particular process based upon field measurements, but other authors generally have either not attempted to collect these data or have been unable to gather the required information to make such an assessment. Their assessments are typically subjective and without data necessary to substantiate that assessment. Even in Reid's (1984) study, logistical, mechanical and other problems limit the amount of data and its accuracy, although he was reasonably certain that about 70% of total erosion resulted from waves, 27% from frost action and 3% from overland flow processes. Slumping was considered a minor process, occurring at one location but modifying a relatively long reach of the shore and a large area of landward bluff sediments.

Reid's (1984) and Sterrett's (1980) studies indicate the present problems with attempting to quantify erosion by process and assess causes. Year-round site specific observations of an interdisciplinary nature are required because of the seasonal nature of some processes, as well as their general episodic nature. Thus, a past erosional process may have actually initiated the present primary erosional process or the erosion and recession that are occurring during the time of the field observations. Procedures need to be developed for accurately determining process importance at a site, although this will be difficult because of the interdependency of single processes and the various shore zone parameters. At present, techniques are lacking to quantitatively correlate the random environmental factors with the erosional processes and the erosional processes with net erosion (Edil and Vallejo 1977). Until site studies of this type are done, predicting erosional or accretionary shoreline changes remains a risky venture fraught with uncertainties.

Simons and Li (1982) attempted to qualitatively evaluate the relative magnitude of forces causing bank erosion in rivers for certain conditions. They used available data, personal experiences, current theories and professional judgment to develop a <u>subjective</u> analysis of the importance of 12 specific erosional processes or variables. Once rated, they calculated the relative magnitude of erosion by each process acting on noncohesive and stratified banks, and then standardized that value to an estimated shear stress acting on a noncohesive bank in a natural river reach. Using the concept of critical shear stress discussed earlier, they then estimated the stability of the material to that stress. Their evaluation of the importance of a particular process (e.g. freeze-thaw) is not always in agreement with those of other authors who essentially used a similar subjective anal-

ysis in considering their relative importance (e.g. Wolman 1959, Hill 1973, Reid 1984).

One approach that may eventually provide a means for assessing the behavior of long reaches of shoreline involves applying the concept of shore zone evolution and slope development. Essentially this concept is that slopes (whether submerged or subaerial) will gradually change with time to approach a configuration in adjustment with the dynamics of the environment at that location. Once that configuration is reached, adjustments are made if the dynamics of the system change. A shore zone within a newly formed reservoir would be the beginning stage while a coastal beach may be an example of the equilibrium form.

This concept is essentially that embodied by Kachugin's (1966) and Hand's (1980) predictive calculations and follows Bruun's (1962) idea of changes in shore zones due to rising water levels. It is also the concept that Brunsden and Kesel (1973) considered in interpreting bluff development along rivers. For example, Brunsden and Kesel (1973) recognized three distinct stages to bluff slope development, each of which is characterized by certain active erosional processes and morphological changes. A similar cyclical change to eroding shore zone bluffs on Lake Michigan was suggested by Quigley et al. (1977) who also found relationships existed for the erosional processes active during each stage of the cycle. Edil and Vallejo (1977) found that eroding bluffs could be subdivided into 3 zones, each with characteristic configurations, erosional processes and changes with time.

The shore zone evolution concept would permit some lumping of the effects of processes, once known, but it would still require that detailed interdisciplinary analyses of eroding bluffs and beaches be undertaken. Landform development models might then be applied (e.g. Craig 1980). This concept needs to be evaluated more fully for analysis of eroding shore zones in reservoirs.

To summarize, no quantitative empirical calculation or theory exists to account for bluff or beach erosion and, in particular, none which considers the complex interaction of properties and processes that determine the occurrence and rate of shore erosion and shoreline recession. The magnitude of importance of a given process at a site is essentially unstudied as are the complex interactions of multiple processes, in part because it is very difficult to separate and isolate a single process in the natural environment. These conclusions apply equally to reservoir, lake and coastal shores.

Clearly, basic research with a truly interdisciplinary approach is needed on a site-specific basis to quantitatively analyze the erosion processes, their interaction and their relative importance in terms of seasonal and long-term changes in beach and bluff zone geometry and position. External forcing variables (such as climate or basin hydrology) must also be considered. Only after comparable site-specific studies are made, can more regional correlations and attempts at developing a predictive capability be done.

Based upon this review, I recommend that site-specific and subsequently regional comparative studies be undertaken. Systematic, quantitative

studies of erosion and recession are needed at multiple sites within northern reservoirs which analyze and interrelate the following:

1. Subaerial and subaqueous erosional, transport and depositional processes within beach and bluff zones,

2. Physical properties (sedimentologic, geotechnical, hydrologic, thermal) of beach and bluff zone materials, including analysis of their seasonal and regional variabilities,

3. Volume of sediment removed by each erosional process (or groups of interrelated processes) and the resulting change in shoreline and bluffline position,

4. Geomorphology and geology of the reservoir basin and surrounding terrain,

5. External forcing variables of the site and region, including climate, regional ground water flow patterns, and reservoir operations.

Studies of this type should be undertaken at many reservoirs across a region, in the hope of identifying regional relationships among erosional processes and factors. Only then could empirical relationships be developed for assessing erodibility and predicting erosion on a reservoir-wide or regional scale. The results of such studies should then be applied to selecting appropriate techniques for reducing or preventing shoreline recession in reservoirs at locations where shore erosion is an unacceptable environmental problem.

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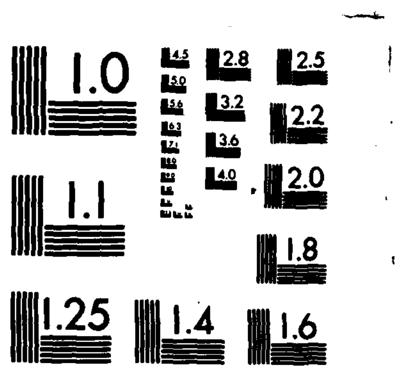

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APPENDIX A: GLOSSARY OF SELECTED BEACH AND WAVE TERMS
(After USACERC 1984)

Amplitude, wave - The magnitude of the displacement of a wave from a mean value. For example, an ocean wave has an amplitude equal to the vertical distance from still-water level to wave crest. For a sinusoidal wave, amplitude is one-half the wave height.

Attenuation - (1) A lessening of the amplitude of a wave with distance from the origin. (2) The decrease of water-particle motion with increasing depth. Particle motion resulting from surface oscillatory waves attenuates rapidly with depth, and practically disappears at a depth equal to a surface wavelength.

Backshore - The zone of the beach profile extending landward from the sloping foreshore (see definition below) to the point of development of vegetation or change in physiography (sea cliff, dune field, etc.).

Beach face - The sloping section of the beach profile below the berm which is normally exposed to the action of the wave swash.

Beach ridge - A nearly continuous mound of beach material that has been shaped by wave or other action. Ridges may occur singly or as a series of approximately parallel deposits.

Beach scarp - An almost vertical escarpment notched into the beach profile by wave erosion. Its height is commonly less than a meter, although higher examples are found.

Berm (beach berm) - A nearly horizontal portion of the beach or backshore formed by the deposition of sediment by the receding waves. Some beaches have more than one berm, while others have none.

Berm crest (berm edge) - The seaward limit of a berm.

Breaker depth - The still-water depth at the point where a wave breaks.

Breaker zone - The portion of the nearshore region in which the waves arriving from offshore reach instability and break. With very simple uniform waves, such as may be generated in a laboratory wave tank, the zone may be reduced to a breaker line. On a wide, flat beach secondary breaker zones may occur in which reformed waves break for a second time.

Buoyancy - The resultant of upward forces, exerted by the water on a submerged or floating body, equal to the weight of the water displaced by this body.

Crest of wave - (1) the highest part of a wave. (2) That part of the wave above still-water level.

Current, littoral - Any current in the littoral zone caused primarily by wave action, e.g., longshore current, rip current.

Current, longshore - The littoral current in the breaker zone moving essentially parallel to the shore, usually generated by waves breaking at an angle to the shoreline.

Eddy - A circular movement of water formed on the side of a main current. Eddies may be created at points where the current passes projecting obstructions or where two adjacent currents flow counter to each other.

Energy coefficient - The ratio of the energy in a wave per unit crest length transmitted forward with the wave at a point in the shallow water to the energy in a wave per unit crest length transmitted forward with the wave in deep water. On refraction diagrams this is equal to the ratio of the distance between a pair of orthogonals at a selected point to the distance between the same pair of orthogonals in deep water.

Foreshore - The sloping portion of the beach profile lying between a berm crest (or in the absence of a berm crest, the upper limit of wave swash at high tide) and the low-water mark of the backrush of the wave swash at low tide. This term is often nearly synonymous with the beach face but is commonly more inclusive, also containing some of the flat portion of the beach profile below the beach face.

Froude number - The dimensionless ratio of the inertial force to the force of gravity for a given fluid flow. It may be given as $F_r = V/Lg$ where V is a characteristic velocity, L is a characteristic length, and g the acceleration of gravity; or as the square root of this number.

Inshore - The zone of the beach profile of variable width extending seaward from the foreshore to just beyond the breaker zone.

Internal waves - Waves that occur within a fluid whose density changes with depth, either abruptly at a sharp surface or discontinuity (an interface) or gradually. Their amplitude is greatest at the density discontinuity or, in the case of a gradual density change, somewhere in the interior of the fluid and not at the free upper surface where the surface waves have their maximum amplitude.

Kinetic energy of waves - In a progressive oscillatory wave, a summation of the energy of motion of the particles within the wave.

Littoral transport - The movement of sediment (littoral drift) in the littoral or nearshore zone by waves and currents. Includes movement parallel (longshore transport) and perpendicular (on-offshore transport) to the shore.

Longshore bar - A ridge of sand running roughly parallel to the shoreline. It may become exposed at low tide. At times there may be a series of such ridges parallel to one another but at different water depths and separated by longshore troughs.

Nearshore zone - In beach terminology an indefinite zone extending seaward from the shoreline well beyond the breaker zone. It defines the area of nearshore currents.

Nearshore current system - The current system caused primarily by wave action in and near the breaker zone, and which consists of four parts: the shoreward mass transport of water; longshore currents; seaward return flow, including rip currents; and the longshore movement of the expanding heads of rip currents.

Offshore - The comparatively flat portion of the beach profile extending seaward from beyond the breaker zone (the inshore) to the edge of the continental shelf. This term is also used to refer to the water and waves seaward of the nearshore zone.

Orbit - In water waves, the path of a water particle affected by the wave motion. In deepwater waves the orbit is nearly circular and in shallow-water waves the orbit is nearly elliptical. In general, the orbits are slightly open in the direction of wave motion, giving rise to mass transport.

Oscillatory wave - A wave in which each individual particle oscillates about a point with little or no permanent change in mean position. The term is commonly applied to progressive oscillatory waves in which only the form advances, the individual particles moving in closed or nearly closed orbits. Distinguished from a wave of translation.

Phase - In surface wave motion, a point in the period to which the wave motion has advanced with respect to a given initial reference point.

Phase inequality - Variations in the tides or tidal currents associated with changes in the phase of the moon in relation to the sun.

Phase velocity - Propagation velocity of an individual wave as opposed to the velocity of a wave group.

Plunge point - (1) For a plunging wave, the point at which the wave curls over and falls. (2) The final breaking point of the waves just before they rush up on the beach.

Potential energy of waves - In a progressive oscillatory wave, the energy resulting from the elevation or depression of the water surface from the undisturbed level.

Refraction (of water waves) - The process by which the direction of a wave moving in shallow water at an angle to the contours is changed. The part of the wave advancing in shallower water moves more slowly than that part still advancing in deeper water, causing the wave crest to bend toward alignment with the underwater contours.

Refraction coefficient - The square root of the ratio of the spacing between adjacent orthogonals in deep water and in shallow water at a selected point. When multiplied by the shoaling factor and a factor for friction and percolation, this becomes the wave height coefficient

or the ratio of the refracted wave height at any point to the deep water wave height. Also the square root of the energy coefficient.

Reynolds number - The dimensionless ratio of the inertial force to the viscous force in fluid motion,

$$R_e = \frac{LV}{\upsilon}$$

where L is a characteristic length, υ the kinematic viscosity, and V a characteristic velocity. The Reynolds number is of importance in the theory of hydrodynamic stability and the origin of turbulence.

Rip current - A strong surface current flowing seaward from the shore. It usually appears as a visible band of agitated water and is the return movement of water piled up on the shore by incoming waves and wind. With the seaward movement concentrated in a limited band its velocity is somewhat accentuated.

Runup - The rush of water up a structure or beach on the breaking of a wave. The amount of runup is the vertical height above stillwater level that the rush of water reaches.

Seiche - A standing wave oscillation of an enclosed water body that continues, pendulum fashion, after the cessation of the originating force, which may have been either seismic or atmospheric or an oscillation of a fluid body in response to a disturbing force having the same frequency as the natural frequency of the fluid system.

Shore - The strip of ground bordering any body of water, whether the ground is rock or loose sediment.

Shoreface - The narrow zone seaward from the low tide shoreline covered by water over which the beach sands and gravels actively oscillate with changing wave conditions.

Shoreline - The intersection of a specified plane of water with the shore or beach (e.g. the high water shoreline would be the intersection of the plane of mean high water with the shore or beach).

Significant wave - A statistical term related to the one-third highest waves of a given wave group and defined by the average of their heights and periods. The composition of the higher waves depends upon the extent to which the lower waves are considered. Experience indicates that a careful observer who attempts to establish the character of the higher waves will record values which approximately fit the definition of the significant wave.

Significant wave height - The average height of the one-third highest waves of a given wave group. Note that the composition of the highest waves depends upon the extent to which the lower waves are considered. In wave record analysis, it is the average height of the highest one-third of a selected number of waves, this number being determined by dividing the time of record by the significant period.

Significant wave period - An arbitrary period generally taken as the period of the one-third highest waves within a given group. Note that the composition of the highest waves depends upon the extent to which the lower waves are considered. In wave record analysis, this is determined as the average period of the most frequently recurring of the larger well-defined waves in the record under study.

Sinusoidal wave - An oscillatory wave having the form of a sinusoid.

Solitary wave - A wave consisting of a single elevation (above the original water surface), its height not necessarily small compared to the depth, and neither followed nor preceded by another elevation or depression of the water surfaces.

Standing wave - A type of wave in which the surface of the water oscillates vertically between fixed points, called nodes, without progression. The points of maximum vertical rise and fall are called antinodes or loops. At the nodes, the underlying water particles exhibit no vertical motion, but maximum horizontal motion. At the antinodes, the underlying water particles have no horizontal motion but maximum vertical motion. They may be the result of two equal progressive wave trains traveling through each other in opposite directions.

Storm surge - A rise above normal water level on a coast due to the action of wind stress on the water surface. Storm surge resulting from a hurricane also includes that rise in level due to atmospheric pressure reduction as well as that due to wind stress.

Surge - The name applied to wave motion with a period intermediate between that of the ordinary wind wave and that of the tide, say from 1/2 to 60 minutes. It is of low height; usually less than 0.3 foot.

Surf zone - The portion of the nearshore region in which borelike translation waves occur following wave breaking. This portion extends from the inner breakers shoreward to the swash zone.

Swash - The rush of water up onto the beach face following the breaking of a wave.

Swash zone - The portion of the nearshore region in which the beach face is alternately covered by the uprush of the wave swash and exposed by the backwash.

Swell - Wind-generated waves that have traveled out of their generating area. Swell characteristically exhibits a more regular and longer period, and has flatter crests than waves within their fetch.

Variability of waves - The variation of heights and periods between individual waves within a wave train. (Wave trains are not composed of waves of equal height and period, but rather of heights and periods which vary in a statistical manner.)

Velocity of waves - The speed at which an individual wave advances.

Wave celerity - Wave speed.

Wave group - A series of waves in which the wave direction, wavelength, and wave height vary only slightly.

Wave height - The vertical distance between a crest and the preceding trough.

Wave hindcasting - The use of historic synoptic wind charts to calculate wave characteristics that probably occurred at some past time.

Wavelength - The horizontal distance between similar points on two successive waves measured perpendicular to the crest.

Wave period - The time for a wave crest to traverse a distance equal to one wavelength. The time for two successive wave crests to pass a fixed point.

Wave of translation - A wave in which the water particles are permanently displaced to a significant degree in the direction of wave travel. Distinguished from an oscillatory wave.

Wave steepness - The ratio of the wave height to the wavelength.

Wave train - A series of waves from the same direction.

Wave trough - The lowest part of a wave form between successive crests. Also that part of a wave below stillwater level.

Wind setup - (1) The vertical rise in the stillwater level on the leeward side of a body of water caused by wind stresses on the surface of the water. (2) The difference in stillwater levels on the windward and the leeward sides of a body of water caused by wind stresses on the surface of the water. (3) Synonymous with wind tide and storm surge. Storm surge is usually reserved for use on the ocean and large bodies of water. Wind setup is usually reserved for use on reservoirs and smaller bodies of water.

Wave spectrum - In ocean wave studies, a graph, table, or mathematical equation showing the distribution of wave energy as a function of wave frequency. The spectrum may be based on observations or theoretical considerations.

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